Constraining Quaternary erosion of the Campine Plateau (NE Belgium) using Bayesian inversion of an in-situ produced $^{10}$Be concentration depth profile

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Abstract. The rate at which low-lying sandy areas in temperate regions, such as the Campine area (NE Belgium), have been eroding during the Quaternary is a matter of debate. Current knowledge on the average pace of landscape evolution in the Campine area is largely based on geological inferences and modern analogies. We applied Bayesian inversion to an in-situ produced $^{10}$Be concentration depth profile in fluvial sand, sampled on top of the Campine Plateau, and inferred the average long-term erosion rate together with three other parameters, i.e., the surface exposure age, inherited $^{10}$Be concentration and sediment bulk density. The inferred erosion rate of $44 \pm 9$ mm/kyr (1σ) is relatively large in comparison with landforms that erode under comparable (palaeo-)climates elsewhere in the world. We evaluate this value in the light of the erodibility of the substrate and sudden base level lowering during the Middle Pleistocene. A denser sampling scheme of a two-nuclide concentration depth profile would allow to include more parameters in the model inversion and further reduce their uncertainty.

1 Introduction

The Campine area is a sandy region which covers part of northeastern Belgium and the southern Netherlands (Fig. 1). It is part of the European sand belt and is drained by rivers that belong to the Scheldt basin. The Campine area roughly coincides with the geological Campine Basin, being the southeastern part of the North Sea Basin. From a geodynamic point of view, the Campine Basin is located in an intermediate position in between the rapidly subsiding Roer Valley Graben in the north, and the uplifting Brabant and Ardennes Massifs in the south (Fig. 2). The Campine Basin has witnessed a long Cenozoic burial history. Post-Rupelian marine and estuarine deposition during the last 30 Ma almost exclusively consists of (glaucosite-rich) sand, up to 300 m thick (Vandenberghe et al., 2004). From the Early to Middle Pleistocene onwards, terrestrial conditions become dominant with deposition of a thick series of fluvial sand and gravel from the rivers Meuse and
Rhine (Figs. 2 and 3). In contrast to what the basinal setting of the Campine region would suggest, distinctive topographic features are preserved in the landscape. An illustrative example is the Campine Plateau, which shows a topographic relief of ca. 50 m relative to the surrounding areas (Fig. 3). To date, quantitative data on the amount and rate of Quaternary erosion of the Campine landscape and the Scheldt basin in general are missing. This stands in contrast to the availability of long-term erosion data from e.g. in-situ produced cosmogenic nuclides for the Meuse and Rhine basins (e.g., Schaller et al., 2001; Dehnert et al., 2011; Rixhon et al., 2011). Such data on catchment-wide erosion rates at multi-millennial timescales are crucial to determine background geological erosion rates to evaluate anthropogenic morphodynamics (Vanacker et al., 2007a), to provide calibration data for landscape evolution models (Bogaart and van Balen, 2000; Foster et al., 2015; Campforts et al., 2016), and to assess the overall stability of the landscape in the framework of long-term management of radioactive waste (Van Geet et al., 2012).

Cosmogenic radionuclides (CRN's) have proven useful to quantify geomorphological processes over timespans covering the last 1 Ma (Schaller et al., 2001). Geomorphological surfaces can be dated by measuring the concentration of in-situ produced cosmogenic nuclides (e.g., \(^{10}\)Be and \(^{26}\)Al) that accumulated at the Earth's surface (Hancock et al., 1999; Hein et al., 2009). As the observed cosmogenic nuclide concentration of a given outcrop is a function of its exposure age and denudation rate, stable (i.e. non-erosing) landforms provide optimal sampling locations for exposure dating (e.g., Rixhon et al., 2011). Most landforms are subject to erosion during exposure, resulting in a decrease of the cosmogenic nuclide concentrations with increasing surface denudation rate (e.g., Dehnert et al., 2011). Braucher et al. (2009) showed that the exposure age (and post-depositional denudation rate) of eroding landforms can be constrained based on a deep (> 1.5 m) depth profile of a single cosmogenic nuclide that is sampled at regular intervals.

The accumulation of in-situ produced cosmogenic nuclides in eroding surfaces is a mathematical function with two parameters that are typically unknown a priori: the exposure age, \(t\) [y], and the post-depositional denudation rate, \(E\) [cm/y]. Unknown model parameters can be estimated by inverse modeling of CRN concentration vs. depth profiles. In this procedure, one iteratively proposes new parameter values until the model fits the observed data up to a pre-specified precision. This has been done for estimating \(t\) and \(E\) by, e.g., Siame et al. (2004) and Braucher et al. (2009). Yet, model and measurement errors together with (measurement) data scarcity introduce considerable uncertainty in optimizing the model parameters. The method proposed by Braucher et al. (2009) accounts to some extent for analytical measurement errors, as it generates several CRN concentration profiles consistent with the (analytical) measurement errors, computes for each model parameter set (e.g., a \(t\)-\(E\) pair) within a grid search the corresponding data misfits and retains the median misfit as the performance associated with a given parameter set. This allows for deriving a robust unique solution but does not consider explicitly model parameter uncertainty and ignores model errors (that is, the model is assumed to be perfect). A more comprehensive quantification of parameter and prediction uncertainty is provided by the Bayesian framework. This approach represents parameter uncertainty by a multivariate “posterior” probability distribution, which is consistent with both prior
information and the (CRN) concentration measurements. In this work, we derive the model parameter posterior probability density function (pdf) by state-of-the-art Markov chain Monte Carlo simulation (see, e.g., Robert and Casella, 2004), accounting under certain assumptions for both measurement and model errors.

The overall objective of this study is to constrain within a Bayesian framework the rate and amount of post-depositional denudation of the headwaters of the Nete catchment. The latter is part of the larger Scheldt basin and is used herein to test and demonstrate the application of Bayesian inversion to CRN concentration vs. depth profiles. The Nete catchment is an interesting test case because the upstream areas of the catchment are located at the northwestern edge of the Campine Plateau, which is covered by coarse gravelly sand from the Early-Middle Pleistocene Rhine and thus constitutes a fluvial terrace from which the depositional age nor the exposure age of the sands is well constrained (Beerten et al., in press).

2 Geomorphological evolution of the Campine area

The post-marine hydrographical evolution of the Campine area started with the final retreat of the sea during the Neogene, as a result of systematic sea level lowering and overall uplift of the bordering areas around the southern North Sea (Miller et al. 2005; Cloething et al. 2007). During the Early Pleistocene, the Meuse followed an eastern course from Liège to the region north of Aachen where it merged with the Rhine (Fig. 2). Tectonic movements along Roer Valley Graben faults, and uplift of the northern margins of the Ardennes-Eifel massif caused the Meuse to breach through its northern interfluve and to follow a completely different course. At the same time, the Rhine shifted its course as well, flowing into the northern part of the Campine area where it merged with the Meuse (Fig. 2). Age control is limited, but this event probably took place around 1 Ma at the earliest, since the confluence area of both rivers was situated in the southeastern part of the Roer Valley Graben prior to 1 Ma, and the area that covers the Campine Plateau today was drained by local 'Belgian' rivers until that time (Westerhoff et al., 2008). Both rivers shifted their course towards a more eastern position by 0.5 Ma at the latest, given the absence of Rhine deposits younger than 0.5 Ma in the depocenter of the Roer Valley Graben (Schokker et al. 2005). The deposits that cover the Campine Plateau are often correlated with the upstream Main Terraces of the Meuse and High Terraces of the Rhine (Paulissen, 1973). Westaway (2001) provides a time window for deposition of Rhine sediments west of the Ville Ridge (High Terraces HT2 and HT3) between 0.5 Ma and 1 Ma. The Rhine sediments on top of the Campine Plateau have been attributed to the Sterksel Formation, which was deposited between ca. 0.6 Ma and 1.1 Ma according to van Balen et al. (2000), and between ca. 0.75 Ma and 1 Ma (post-Jaramillo Early-Pleistocene) according to Gullentops et al., (2001). Recently, deposits from the High Terraces of the Rhine between Bonn (Germany) and Venlo (the Netherlands) have been dated to 750 ± 250 ka and 740 ± 210 ka using in-situ produced cosmogenic radionuclides (Dehnert et al., 2011). Similarly, Meuse terrace deposits in the Liège area (Romont, Belgium) that are generally assumed to correspond with the series of Main Terrace deposits, have been dated to 725 ± 120 ka using the same technique (Rixhon et al., 2011).
During the Middle Pleistocene, the hydrography of northern Belgium drastically changed due to the 'opening' of the English Channel (Vandenberghe and De Smedt 1979; Fig. 2). Various studies (Gibbard 2007; Gupta et al. 2007; Toucanne et al. 2009) link the opening of the English Channel to the catastrophic drainage of a large proglacial lake during marine isotope stage 12 (MIS 12), approximately 450 ka ago (Elsterian). The 450 ka event triggered the formation of a buried palaeo-channel system known as the Flemish Valley, with extensions towards the south and the east (Tavernier and De Moor, 1974). The Nete catchment is generally considered as the eastern extension of the Flemish Valley.

At present, the Campine Plateau is a landform that markedly stands out with respect to its surroundings. It is a fluvial terrace covered by coarse gravelly Meuse deposits in the south(east) and sandy Rhine deposits in the north (Fig. 3). The sediments have proven to feature a periglacial palaeoenvironment and were deposited by braided rivers (Paulissen, 1973 and 1983). The Campine Plateau can be considered as a classical case of relief inversion (Fig. 4). Undoubtedly, the area west of the Campine Plateau experienced prolonged phases of erosion and denudation after the Rhine had left the region, around 0.5 Ma at the latest (Fig. 4b; Beerten et al., in press).

### 3 Material and methods

#### 3.1 Cosmogenic radionuclide profiling

Cosmogenic radionuclides (CRN) allow us to quantify geomorphological processes over timespans covering the last 1 Ma. In this study, we use the concentration vs. depth profile of a single in-situ produced CRN ($^\text{10}\text{Be}$) to constrain the post-depositional denudation rate, $E$ [cm/y] of the fluvial terrace. The accumulation of CRN, $N_{\text{total}}(z,t)$ [at/g], in an eroding surface can be described by a mathematical function composed of two terms that represent the inherited CRN concentration of the fluvial sediment, $N_{\text{inh}}(z)$, and the post-depositional production of CRN, $N_{\text{exp}}(z)$:

$$N_{\text{total}}(z,t) = N_{\text{inh}}(z) + \sum_i \frac{P_i(z)}{\lambda_i + \rho \Lambda_i} e^{-\left(\frac{z_0 - E t}{\rho \Lambda_i}\right)/\lambda_i} \left(1 - e^{-\left(\frac{z - z_0}{\rho \Lambda_i}\right)/\lambda_i}\right)$$  (1)

with $t$ [y] the exposure age, $\lambda$ [1/y] the decay constant ($\lambda = \ln 2 / t_{1/2}$), $z_0$ the initial shielding depth ($z_0 = E \times t$), $\rho$ [g/cm$^3$] the density of the overlying material, $\Lambda_i$ [g/cm$^2$] the attenuation length. The production rate of CRN, $P_i(z)$ [at/g/y], is a function of the depth, $z$ [cm], below the surface as:

$$P_i(z) = P_i(0) e^{-\frac{z_0}{\Lambda_i}}$$  (2)

The subscript ‘i’ indicates the different production pathways of in-situ produced $^\text{10}\text{Be}$ via spallation, muon capture and fast muons following Dunai (2010). In this study, we set the relative contribution of neutrons and muons (negative and fast) to
the total $^{10}$Be production at resp. 97.85, 1.50 and 0.65%, and set the effective attenuation length at resp. 150, 1500 and 5300 g/cm² following Braucher et al. (2003). Production rates were scaled following Dunai (2001) with a sea level high latitude production rate of (4.5 ± 0.3) at/g/y. This production rate is consistent with the value that was used earlier on by Rixhon et al. (2011) for dating terraces of the Ardennian rivers. A half-life of $(1.387 ± 0.012) \times 10^6$ y was used for $^{10}$Be following Chmeleff et al. (2010). CosmoCalc add-in for Excel was used to calculate the scaling factors. Given the flat topography of the Campine Plateau, topographic shielding was negligible and therefore not corrected for (Norton and Vanacker, 2009).

### 3.2 Sampling and analytical methods

The depth profile was sampled in a sand pit (SRC-Sibelco NV) on the northwestern edge of the Campine Plateau (Fig. 4a and b). The altitude of the sampling spot is ca. 47 m (Tweede Algemene Waterpassing), while the crest of the plateau further east reaches an altitude of ca. 48 m. The almost 4 m thick sequence is composed of medium-grained quartz-rich fluvial sands, overlain by a thin layer (35 cm thick) of fine-grained aeolian sand (Fig. 5). Detailed grain-size characteristics of the fluvial sand are given in Fig. 6. Note that sample depth is given with reference to the top of the fluvial sands. The lowermost unit A consists of medium sand with mode and median in the range between 250-500 µm, while a significant portion of grains coarser than 500 µm is present. Unit B is finer with a median grain size of ca. 250 µm and virtually no coarse sand (i.e., > 500 µm). Unit C consists of coarse sand (median grain size more than 500 µm) with a significant amount of fine gravel fragments. The next unit (E) is the finest unit of the sequence, with mode and median below 250 µm. Sediments from unit F are generally finer than those of units A and C, but coarser than those from units B and D. Mode and median are in the range between 250-500 µm. Finally, unit G represents a thin layer of fine sand, interpreted as Late Pleistocene aeolian deposits. Based on the grain size analysis, the fluvial sequence can be divided into several fining-upward sequences, where samples MHR-II-06 and MHR-II-04 are taken in much finer sand beds compared to the other samples.

From the depth profile, ten samples were collected for CRN analysis at depths ranging from 10 to 320 cm below the fluvial-aeolian contact, from which 9 were analysed. Samples were more or less evenly spread out over the sequence, although the sampling density was higher towards the top (Table 2). Samples were taken as bulk samples of 1.5 kg, over a depth interval of 10 cm. Samples were sieved, and the 500-1000 µm grain size fraction was used for sample preparation, except for the fine-grained sand samples MHR-II-04 and MHR-II-06 where the 250-500 µm fraction had to be used.

Samples were prepared at the University of Louvain Cosmogenic Isotope Laboratory (Louvain-la-Neuve). In-situ produced $^{10}$Be was extracted from purified quartz using standard separation methods described in von Blanckenburg et al. (1996) and Vanacker et al. (2007b). Two blanks were processed with the nine samples. Approximately 200 µg of $^9$Be carrier was added to blanks and samples containing 30 to 35 g pure quartz. $^{10}$Be/$^9$Be ratios were measured in BeO targets with accelerator mass spectrometry on the 0.6 MV Tandy at ETH Zurich (Kubik and Christl, 2010). The ratios were normalized to the ETH in-house secondary standard S2007N with a nominal value of $^{10}$Be/$^9$Be of $28.1 \times 10^{-12}$ (Kubik and Christl, 2010) which is in agreement with a half-life of 1.387 Myr (Chmeleff et al., 2010). Samples are corrected for the number of $^{10}$Be atoms in their
associated blanks. The analytical uncertainties on the $^{10}$Be/$^{9}$Be ratios of sample and blank are then propagated into the $1\sigma$ analytical uncertainty for nuclide concentrations.

### 3.3 Bayesian inference

#### 3.3.1 Inverse problem

To acknowledge that measurements and modelling errors are inevitable, the inverse problem is commonly represented by the stochastic relationship given by

$$F(x) = d + e$$  \hspace{1cm} (3)

where $F$ is a deterministic, error-free forward model that expresses the relation between the uncertain parameters, $x$, and the measurement data, $d$, and the noise term, $e$, lumps measurement and model errors.

Inversions were performed within a Bayesian framework, which treats the unknown model parameters $x$ as random variables with posterior probability density function (pdf), $p(x|d)$, given by

$$p(x|d) = \frac{p(d|x)p(x)}{p(d)} \propto L(x|d)p(x)$$  \hspace{1cm} (4)

where $p(x)$ denotes the prior distribution of $x$ and $L(x|d) \equiv p(d|x)$ signifies the likelihood function of $x$. The normalization factor $p(d) = \int p(d|x)p(x)dx$ is obtained from numerical integration over the parameter space so that $p(x|d)$ scales to unity. The quantity $p(d)$ is generally difficult to estimate in practice but is not required for parameter inference. In the remainder of this study, we will focus on the unnormalized posterior $p(x|d) \propto L(x|d)p(x)$. For numerical stability, it is often preferable to work with the log-likelihood function, $\ell(x|d)$, instead of $L(x|d)$. If we assume the error $e$ to be normally distributed, uncorrelated and with unknown constant variance, $\sigma^2$, the log-likelihood function can be written as

$$\ell(x|d) = -\frac{N_d}{2} \ln(2\pi) - \frac{N_d}{2} \ln(\sigma^2) - \frac{1}{2\sigma^2} \sum_{i=1}^{N_d} [d_i - F_i(x)]^2$$  \hspace{1cm} (5)

where $\sigma^2$ can be fixed beforehand or sampled jointly with the other model parameters $x$. It is worth noting that by fixing $\sigma$ to a known measurement error, one implicitly assumes that the model is able to describe the observed system up to the observation errors. This might not be realistic in environmental modelling, where models are always fairly simplified descriptions of a more complex reality. In this work, we therefore jointly infer $\sigma$ with $x$. 

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3.3.2 Markov chain Monte Carlo sampling

The main goal of the inference is to estimate the posterior parameter distribution of the model parameters, \( p(\mathbf{x}|\mathbf{d}) \). As an exact analytical solution for \( p(\mathbf{x}|\mathbf{d}) \) is not available, we resort to Markov chain Monte Carlo (MCMC) simulation to generate samples from this distribution. The basis of this technique is a Markov chain that generates a random walk through the search space and iteratively finds parameter sets with stable frequencies stemming from the posterior pdf of the model parameters (see, e.g., Robert and Casella, 2004, for a comprehensive overview of MCMC simulation). In practice, the MCMC sampling efficiency strongly depends on the assumed proposal distribution used to generate transitions in the Markov chain. In this work, the state-of-the-art DREAMzs (ter Braak and Vrugt, 2008; Vrugt et al., 2009; Laloy and Vrugt, 2012) algorithm is used to generate posterior samples. A detailed description of this sampling scheme including convergence proof can be found in the literature cited and is thus not reproduced herein.

To constrain the post-depositional erosion rate, \( E \), of the sampling site on the Campine Plateau, we apply Bayesian inference on the mathematical expression of the CRN concentration vs. depth profile (Eq. (1)). Details on the model parameter settings are given above. The following marginal prior distributions were defined for the sampled five model parameters, based on the current geological knowledge of the region, as well as the measured \(^{10}\text{Be} \) data (Table 1 and 2). For 3 out of 5 parameters, i.e., post-depositional erosion (\( E \)), exposure age (\( t \)) and bulk density (\( \rho \)), we specified truncated Gaussian prior distributions using mid-range points of the adopted interval as means and 0.4 times the interval range as standard deviations. Our justification for this choice is as follows. On the one hand, the prior distribution is sufficiently vague to avoid over-constraining the search (mid-range prior probability is only about twice that of the boundary values), and, on the other hand, it discourages the search to pick up values close to the boundaries as they are considered to be (a priori) less likely. The exposure age of the terrace, \( t \) [yr], was allowed to vary between 0.5 Ma and 1 Ma, based on the presumed burial age of the Rhine sands covering the Campine Plateau at that location from geological evidence (see Section 2). This results in a prior mean of 0.75 Ma, and a standard deviation of 0.2 Ma. These estimates are fully consistent with CRN burial ages of 750 ± 250 ka and 740 ± 210 ka that were established for the High Terrace deposits from the Rhine in Germany and the Netherlands (Dehnert et al., 2011). The denudation rate (\( E \)) prior distribution was translated into a mean of 30 m/Myr and standard deviation of 24 m/Myr, with upper and lower bounds of 60 m/Myr and 0 m/Myr, respectively. The maximum value is based on geomorphological evidence presented in Fig. 4b, using the altitude difference between the Campine Plateau and the adjacent Kleine Nete floodplain, and taking into account the youngest possible age for the Rhine sediments. The minimum value is set to equal the null-hypothesis, to encompass the scenario of the Campine Plateau being a residual relief due to erosion-resistance of the substrate. The mean prior bulk density of the studied fluvial sediments was set to 1.65 g/cm\(^3\), which is consistent with the average value of upper Neogene and Quaternary sediments in the region (Beerten et al., 2010). The full range of the prior for bulk density was set between 1.5-1.8 g/cm\(^3\), to allow for some variation in grain size, sorting and sediment packing. Bulk densities below 1.5 g/cm\(^3\) are typical for clay-rich and or organic-rich soils, while densities in excess
of 1.8 g/cm\(^3\) are typical for gravel-bearing sediments. However, none of these lithologies have been observed in such quantities so as to justify a wider prior range for the bulk density.

In contrast to the other model parameters, the prior distribution for the inherited \(^{10}\)Be concentration, \(N_{\text{inh}}\), was assumed to be uniform. This is because solutions for \(N(z,t)\) in Eq. (1) always converge to \(N_{\text{inh}}\) as \(z\) tends to infinity while no \(N_{\text{inh}}\) measurements are available at \(z\) larger than 3.5 m. The \(N_{\text{inh}}\) parameter was therefore allowed to vary uniformly between two extreme values. The maximum value of \(9 \times 10^4\) at/g is consistent with the highest concentration measured in the profile (Table 2; the inherited concentration cannot be higher than this value). The lower bound was somewhat arbitrarily set at \(1 \times 10^4\) at/g, given the fact that zero inheritance is considered to be very unlikely.

Lastly, a so-called Jeffreys (1946) prior of the form \(p(\sigma) \propto 1/\sigma\) is used for \(\sigma\). This classical choice basically means that one wants to achieve \(\sigma\) values that are both as small as possible and consistent with the data misfit.

### 3.3.3 Predictive uncertainty intervals

A 95% uncertainty interval for the simulated CRN concentrations can be calculated by drawing parameter sets, \([x, \sigma]\), from \(p(x|d)\) and removing the 2.5% largest and lowest values from the associated set of \(F(x)\) responses. We call the so-derived 95% uncertainty interval a “95% uncertainty interval due to parameter uncertainty”. Because of model inadequacy (that is, model errors), this interval may not bracket 95% of the observations though. Consistently with the assumptions underlying Eq. (5), we thus compute a “95% total uncertainty interval” by adding to each \(F(x)\) a random noise \(e \propto N(0, \delta)\), with \(\delta = \sigma - \sigma_m\) where \(\sigma_m\) is a fixed measurement error which we take as the maximum analytic measurement error of our dataset, that is, \(\sigma_m = 2000\) atoms/g.

### 4 Results

#### 4.1 CRN profiling technique

Table 2 lists sample input parameters, and measured \(^{10}\)Be concentrations with analytical errors. The depth is given below the contact with the overlying aeolian sand cover. In general, there is a clear decrease in \(^{10}\)Be concentration with depth, except for two samples (-04 and -06) which clearly contain higher CRN concentrations (Fig. 6). It is striking that the CRN concentrations are consistently higher for the two samples were the finer (250-500 µm) grain size fraction was analysed. Grain size-dependent \(^{10}\)Be concentrations can point to differences in geomorphological process rates in the regions of sediment provenance as suggested by Carretier et al. (2015). Alternatively, the negative relation between in-situ produced CRN and grain size might also result from non-stationary sedimentation rates, where samples from the fine-grained layers accumulated CRN during the final stage of the sedimentation cycle prior to a phase of non-deposition and/or steady state. Apart from the -04 and -06 samples, the CRN concentrations decrease non-linearly with depth, from \((1.5 \pm 0.02) \times 10^5\) at/g...
at 10 cm to \((9.0 \pm 0.2) \times 10^4\) at/g at 320 cm. If we assume that the Campine Plateau is an erosion-resistant landform with negligibly low erosion rates \((E \approx 0 \text{ cm/yr})\), the difference between \(N_{\text{total}}(z)\) and \(N_{\text{inh}}\) gives the \(N_{\text{exp}}(z)\) or the concentration of cosmogenic nuclides that is produced at depth \(z\) after deposition of the Rhine sands. The apparent exposure age of the surface, \(t\), can then be reconstructed following Eq. (1). By doing so, we obtain an apparent exposure age of \(21.5 \pm 1.5\) ka, which is in strong contradiction with chronosтратigraphical age estimates of the fluvial deposits that cover the Campine Plateau that range between 0.5 and 1 Ma (see Section 2 and 3.3.2). We advocate that post-depositional erosion has strongly altered the \(^{10}\text{Be}\) signature of the upper layers of the Rhine sediments at the study site.

### 4.2 Posterior parameter distribution

The marginal posterior distributions of the 5 sampled parameters (including the standard deviation of the residual errors) are depicted in Fig. 7. The \(t\) distribution shows a weakly expressed mode around the prior mean value (Fig. 7a) whereas no significant correlation is observed between \(t\) and the other model parameters except for \(N_{\text{inh}}\) with which a linear correlation of -0.13 is observed (Fig. 8a-d-f and Table 3). This means that the measurement data do not contain enough information to resolve the exposure age. In other words, depending on the other model parameter values, the model can always fit the data no matter the value of \(t\) (within its prior range of 0.5 Ma - 1 Ma). The erosion rate, \(E\), is much better resolved with a posterior mode around \(4.4 \times 10^3\) cm/yr which is equal to 44 mm/kyr or m/kyr (Fig. 7b). The posterior \(E\) uncertainty is relatively small, being 9 cm/yr (1\(\sigma\)) and 15 cm/yr (2\(\sigma\)). A large posterior linear correlation coefficient of 0.71 is found between \(E\) and \(N_{\text{inh}}\) (Fig. 8c and Table 3). This suggests that one could try to fix one and only infer the other. The posterior bulk density distribution does not deviate much from its prior distribution (Fig. 7d), and shows a weak (negative) correlation with \(E\) (Fig. 8b and Table 3). The \(N_{\text{inh}}\) parameter shows a clear mode around \(6.5 \times 10^4\) atoms/g, which is lower than the lowest measured value in the profile of about \(9.09 \times 10^4\) atoms/g (Fig. 7d; Table 2). As mentioned already, a strong dependence with \(E\) was found (Fig. 8c and Table 3). The standard deviation of the residual errors, \(\sigma\), shows a log-normal marginal posterior with a clear mode around \(1 \times 10^4\) atoms/g (Fig. 7e). This is consistent with the achieved root mean square error (RMSE) values between measured and simulated \(^{10}\text{Be}\) that follow a similar distribution with a mode about \(1 \times 10^4\) atoms/g (not shown). It is interesting to note that with a range between \(1.4 \times 10^3\) and \(2 \times 10^3\) atoms/g, the analytical measurement errors are more than 5 times smaller than the values taken by \(\sigma\). This nicely illustrates the effect of model errors. If the model would have been perfect, the achieved RMSE values and \(\sigma\) distribution should indeed have been within this measurement error range of \(1.4 \times 10^3\) to \(2 \times 10^3\) atoms/g.

Fig. 9 presents the 95% predictive uncertainty interval due to parameter uncertainty (dark gray area) together with the 95% total predictive uncertainty interval (light gray area). Indeed, the 95% uncertainty interval due to parameter uncertainty brackets only 6 observations out of 7, which is less that what would be expected from a 95% confidence level. This gap is likely caused by model inadequacy. The derived 95% total predictive uncertainty interval is much larger and encapsulates all of the 7 observed data.
Discussion

Five parameters of the CRN model were considered to be uncertain within the MCMC inversion: $E$, $t$, $N_{\text{inh}}$, $\rho$ and $\sigma$. With only seven measurement data, the chosen marginal prior distributions for these parameters are obviously strongly influencing the sampled posterior distribution. Moreover, as stated earlier, we used relatively flat distributions with wide ranges such that the search towards the maximum a posteriori (MAP) prediction is not much constrained by the prior. This resulted in a poor resolution for the $t$ and $\rho$ parameters (Fig. 7a and 7c) for which marginal prior and posterior distributions are quite similar. In contrast, the inherited $^{10}$Be concentration ($N_{\text{inh}}$) appears to be a well-resolved parameter. This is an important observation, because for eroding surfaces such as the one considered in this study, the inherited concentration contributes significantly to the total $^{10}$Be concentration. The target parameter in this study, that is, the erosion rate ($E$), appears to be well-resolved as well, showing a relatively tight posterior distribution that is shifted towards higher mean values compared to the prior (Fig. 7b). In the following, the posterior mean of the $E$ distribution, together with a one sigma uncertainty level, i.e. $44 \pm 9$ m/Myr or mm/kyr, are used to discuss the inferred erosion rate.

In Fig. 10, the pooled erosion rate data for outcrops, as published by Portenga and Bierman (2011) are shown, together with the mean erosion rate obtained in the present study. The global erosion data are based on surface samples (thickness ranging between 0.5 cm and 8 cm) from a variety of bedrocks, including igneous, metamorphic and sedimentary rocks, and various climato-tectonic settings. Generally speaking, outcrop erosion rates from Portenga and Bierman (2011) tend to be lower than the one determined for the headwaters of the Nete catchment, with a few exceptions. Since their data set is entirely based on bedrock samples, we argue that the differences in erosion rate principally reflect differences in erosion resistance of the substrate, i.e., consolidated rock versus unconsolidated sediment. Furthermore, in a western European context, the erosion rate that we report here seems to be fairly high for a fluvial terrace. For comparison, the Meuse younger Main Terrace (YMT) near Liège does not show any signs of post-depositional erosion following Rixhon et al. (2011). Probably, the coarse-grained and slightly consolidated nature of the Meuse sediment can be put forward as an explanation. Similarly, Dehnert et al. (2011) reported that the High Terraces of the Rhine in Germany and the Netherlands were eroded by only 1-3 m, and that the loess cover presumably protected the Rhine sands from significant erosion soon after deposition.

An alternative explanation for the relatively high erosion rate found in the current study may be the proximity of the North Sea, and its changes in base level during sea level lowstands. In contrast to the Meuse and Rhine, the Scheldt basin, to which the studied Rhine terrace belongs today, developed in response to the sudden base level lowering as a result of the opening of the English Channel, ca. 450 ka ago (see Section 2). An important feature of the Scheldt basin is the Flemish Valley, a buried river system, which the Nete basin is an eastern extension from. We advocate that the sudden base level lowering may have caused a regressive erosion wave penetrating into the hinterland, shaping the Flemish Valley and its eastern extension, i.e. the Nete catchment, and cause increased erosion rates in this distal part of the Scheldt basin. An important issue to tackle is to which extent the currently obtained erosion rate can be used to infer the total amount of erosion at the site. Mathematically, this translates into $E \times t$, and would roughly yield a value of between ca. 22 m (for $t = ...
500 kyr) and 44 m (for t = 1 Myr). Adding up with the current thickness of fluvial sediment at the site (i.e., 5-10 m) this would mean that the original thickness of Rhine sediment would have been between ca. 27-54 m. Since this would exceed the thickness of Rhine sediment in the southwestern part of the Roer Valley Graben (Beerten, 2006), we consider this hypothesis to be false. Instead, we suggest that erosion started later, presumably after opening of the English Channel around 450 ka. The $^{10}$Be data thus would reflect the arrival of the headward erosion wave in the distal part of the Nete catchment, following the opening of the English Channel, and the reorganisation of the hydrographical network. On the one hand this implies that the total amount of erosion is less than 22 m, according to the formula $E \times t$. On the other hand, at least 5-10 metres of extra sediment would have been needed to completely shield the sampled sediment from cosmogenic radionuclide production before the start of erosion, sometime after 450 ka.

In any case, there are several implications for the regional evolution of the landscape attached to the observation of sediment removal on top of the Campine Plateau. Firstly, it would mean that the total amount of Rhine sediment would have been larger than what can be observed today in the quarries (Fig. 4); this should be taken into account when correlating Rhine sediment from the Campine Plateau with that in the Roer Valley Graben. Secondly, post-depositional fault movement along (segments of) the Feldbiss fault as derived from the stratigraphy of Rhine sands should be considered as a minimum (Fig. 3). Thirdly, the amount of post-depositional erosion west of the plateau (Fig. 2) as can be observed from present-day altitude differences (Campine Plateau vs. Nete valley; Fig. 4b) should be regarded as a minimum erosion value.

In future work, the aim is to consider more parameters in the MCMC inversion when new and more densely sampled profiles become available. Moreover, in an attempt to account for differences in geomorphological process rates in the regions of sediment provenance or non-stationary sedimentation rates, resulting in grain-size dependent $^{10}$Be concentrations, the MCMC simulation could be combined with a distributed numerical forward modelling approach instead of the currently used analytical solution.

### 6 Conclusion

In accordance with previous studies, we conclude and confirm that a single $^{10}$Be concentration depth profile can generate meaningful and reliable information on the denudation rates of a given landform, even when several parameters are highly uncertain. An MCMC inverse modeling procedure leads to a maximum a posteriori post-depositional erosion rate of ca. 44 ± 9 m/Myr (mm/kyr) for the studied fluvial terrace of the Rhine which today belongs to the Nete catchment. This derived value of post-depositional erosion rate is fairly high compared to published outcrop erosion rate data in the Meuse and Rhine catchment, and elsewhere in the world. It is inferred that the unconsolidated nature of the studied sediment, the absence of a protecting cover (such as loess) and possibly also headward erosion in response to sudden base level lowering around 450 ka are put forward as a possible explanation. We conclude that MCMC modelling of $^{10}$Be concentration depth profiles is a powerful tool to determine rates of landscape evolution and earth surface processes and to assess the associated uncertainties.
7 Code availability

A MATLAB code of the Bayesian inversion used in this study is available upon request to the second author.

8 Author contributions

K. Beerten designed the study, performed the sampling and prepared the manuscript with contributions from the second and third authors. E. Laloy implemented the model and performed the simulations. V. Vanacker prepared the samples, interpreted and processed the raw data and performed initial calculations. B. Rogiers co-designed the study and L. Wouters supervised it.

9 Acknowledgements

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10 References


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Figure Captions

Figure 1 – Location of the Campine region within Europe and the European Sand Belt.

Figure 2 – Structural map of northwestern Europe showing the Roer Valley Graben faults and the Brabant and Rhenohercynian (Ardennes) massifs superimposed on a Digital Terrain Model (DTM) of northwestern Europe (GTOPO30; data available from the U.S. Geological Survey), with indication of large rivers (http://www.eea.europa.eu/data-and-maps/data/wise-large-rivers-and-large-lakes) and location of the Scheldt basin, (dashed line), the Nete catchment (dotted line) and the Flemish Valley (solid line). The general palaeohydrography of the Meuse and Rhine between 0.5 Ma and 1.0 Ma is shown in coloured lines. Headward erosion as an explanation for the development of the Nete catchment is indicated with a yellow arrow.

Figure 3 – DTM of the Campine Plateau (Digitaal Hoogtemodel Vlaanderen II, DTM, raster, 1 m) and the extent of Rhine deposits in the study area (shading; based on Beerten (2005) and Deckers et al. (2014)).

Figure 4 – (a): Detailed DTM of the study area (Digitaal Hoogtemodel Vlaanderen II, DTM, raster, 1 m), with indication of the sampling location (white arrow). Note the regularly shaped sand quarries south of profile line A-A' which appear as depressions on the DTM. (b): Topographic cross-section according to the profile line (A-A') shown in (a). The sampling location is schematically shown as a grey rectangle.

Figure 5 – Photograph of the sampled profile with indication of sampling points, field codes, lithological units (A-G) and approximate profile depth.

Figure 6 – $^{10}$Be concentration profile and results of the grain size analyses. Note that the elevated $^{10}$Be concentrations belong to samples that were analysed using a smaller grain size fraction than the other samples (i.e., 250-500 µm instead of 500-1000 µm).

Figure 7 – Posterior distributions of the six parameters. (a): exposure age. (b): erosion rate. (c): bulk density. (d): inherited $^{10}$Be concentration. (e): standard deviation of the residual errors.

Figure 8 – Selected correlation plots for the five parameters. Note that $N_{inh}$ and $E$ show the strongest correlation.
Figure 9 – 95% predictive uncertainty interval due to parameter uncertainty and the 95% total predictive uncertainty interval (dark gray area). Depth is given in cm below the contact with the overlying aeolian sand cover.

Figure 10 – Frequency distribution of the pooled outcrop erosion rates obtained from Portenga and Bierman (2011), with indication of the mean value obtained for the Nete catchment (this study).
Table 1 - Prior distributions for four parameters: erosion rate, exposure age, bulk density and inherited $^{10}$Be concentration.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Unit</th>
<th>Distribution</th>
<th>Min-max</th>
<th>Mean</th>
<th>St. dev.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Erosion rate</td>
<td>$E$</td>
<td>m/Myr or mm/kyr</td>
<td>Truncated Gaussian</td>
<td>0 – 60</td>
<td>30</td>
<td>24</td>
</tr>
<tr>
<td>Exposure age</td>
<td>$t$</td>
<td>Ma</td>
<td>Truncated Gaussian</td>
<td>0.5 – 1.0</td>
<td>0.75</td>
<td>0.2</td>
</tr>
<tr>
<td>Bulk density</td>
<td>$\rho$</td>
<td>g/cm$^3$</td>
<td>Truncated Gaussian</td>
<td>1.5 – 1.8</td>
<td>1.65</td>
<td>0.12</td>
</tr>
<tr>
<td>Inherited $^{10}$Be</td>
<td>$N_{inh}$</td>
<td>$10^4$ atoms/g</td>
<td>Uniform</td>
<td>1 – 9</td>
<td>n/a</td>
<td>n/a</td>
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</table>

Table 2 - Analytical results from the in-situ produced 10Be analysis. The depth profile is located at 50.95°N and 5.63°W at an altitude of 47 m. A SLHL (sea-level high-altitude) production rate of 4.5 ± 0.3 at/g/y was used in this study. We refer to the text for more information on the methodology used.

<table>
<thead>
<tr>
<th>Sample label</th>
<th>Sample field code</th>
<th>Relative depth $z_{fluv}$ (cm)*</th>
<th>Quartz (g)</th>
<th>Be carrier (mg)</th>
<th>$^{10}$Be/Be ($\times 10^{-12}$)</th>
<th>$^{10}$Be conc ($\times 10^5$at/g qtz)</th>
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<tbody>
<tr>
<td>TB1204</td>
<td>BE-MHR-II-00</td>
<td>10</td>
<td>34.406</td>
<td>0.208</td>
<td>0.388 ± 0.016</td>
<td>1,537 ± 0,019</td>
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<tr>
<td>TB1205</td>
<td>BE-MHR-II-01</td>
<td>40</td>
<td>33.535</td>
<td>0.207</td>
<td>0.329 ± 0.016</td>
<td>1,328 ± 0,017</td>
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<tr>
<td>TB1206</td>
<td>BE-MHR-II-02</td>
<td>60</td>
<td>33.370</td>
<td>0.207</td>
<td>0.252 ± 0.016</td>
<td>1,015 ± 0,015</td>
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<tr>
<td>TB1207</td>
<td>BE-MHR-II-03</td>
<td>80</td>
<td>34.467</td>
<td>0.207</td>
<td>0.318 ± 0.016</td>
<td>1,245 ± 0,017</td>
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<tr>
<td>TB1940</td>
<td>BE-MHR-II-04</td>
<td>120</td>
<td>23.478</td>
<td>0.164</td>
<td>0.521 ± 0.019</td>
<td>2,397 ± 0,029</td>
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<td>TB1208</td>
<td>BE-MHR-II-05</td>
<td>160</td>
<td>34.620</td>
<td>0.207</td>
<td>0.231 ± 0.015</td>
<td>0,898 ± 0,014</td>
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<tr>
<td>TB1944</td>
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<td>0,709 ± 0,043</td>
<td>3,272 ± 0,037</td>
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<tr>
<td>TB1209</td>
<td>BE-MHR-II-07</td>
<td>240</td>
<td>34.186</td>
<td>0.207</td>
<td>0,251 ± 0,014</td>
<td>0,987 ± 0,015</td>
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<tr>
<td>TB1210</td>
<td>BE-MHR-II-09</td>
<td>320</td>
<td>33.663</td>
<td>0.207</td>
<td>0,229 ± 0,014</td>
<td>0,909 ± 0,015</td>
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<tr>
<td>TB1211</td>
<td>BE-BLANK-01</td>
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<td>0.000</td>
<td>0.207</td>
<td>0,011 ± 0,0006</td>
<td>0,0041 ± 0,0009</td>
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<tr>
<td>TB1941</td>
<td>BE-BLANK-02</td>
<td>NA</td>
<td>0.000</td>
<td>0.164</td>
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</table>

*The relative depth $z_{fluv}$ is given as depth below the contact with the overlying aeolian sand cover.
Table 3 – Posterior linear correlation coefficients between the 5 sampled variables.

<table>
<thead>
<tr>
<th>Sampled parameter</th>
<th>E</th>
<th>t</th>
<th>ρ</th>
<th>N_{inh}</th>
<th>σ</th>
</tr>
</thead>
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<tr>
<td>E</td>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>t</td>
<td>0.017</td>
<td>1</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ρ</td>
<td>-0.221</td>
<td>-0.002</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N_{inh}</td>
<td>0.709</td>
<td>-0.139</td>
<td>0.197</td>
<td>1</td>
<td></td>
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<tr>
<td>σ</td>
<td>-0.110</td>
<td>0.007</td>
<td>-0.082</td>
<td>-0.145</td>
<td>1</td>
</tr>
</tbody>
</table>
Figure 01
Figure 02

- National boundaries
- Rivers
- Nete catchment
- Scheldt basin
- Flemish Valley
- Roer Valley Graben faults
- Brabant Massif
- Rhenohercynian Massif

Elevation (m a.s.l.):
- 0
- 20
- 40
- 60
- 80
- 100
- 120
- 140
- 160
- 180
- >200

River palaeogeography:
- Rhine 0.5-1.0 Ma
- Meuse > 1.0 Ma
- Meuse 0.5-1.0 Ma

headward erosion

Fig. 3
Figure 03
Figure 04

(a) Map showing the location of the Campine Plateau and the Rhine sands. The Kleine Nete valley is indicated as well.

(b) Line profile along the profile line A’ showing the 30 m denudation and the Plio-Pleistocene substrate buried under thin aeolian deposits and shallow floodplain deposits. The Campine Plateau and Rhine sands are also indicated.
Figure 05
Figure 06
Figure 07
Figure 08
Figure 09

The figure shows a graph with depth in cm on the y-axis and in-situ $^{10}$Be [atoms/g] on the x-axis. The graph includes data points marked with red crosses, representing measurements. The shaded areas represent uncertainty, with lighter shading for total uncertainty and darker shading for 95% parameter uncertainty. The line with a black arrow indicates the MAP prediction.
Figure 10

This study