CE-DYNAM (v1), a spatially explicit, process-based carbon erosion scheme for the use in Earth system models

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Abstract.

Soil erosion by rainfall and runoff is an important process behind the redistribution of soil organic carbon (SOC) over land, hereby impacting the exchange of carbon (C) between land, atmosphere and rivers. However, the net role of soil erosion in the global C cycle is still unclear as it involves small-scale SOC removal, transport and re-deposition processes that can only be addressed over selected small regions with measurements and models. This leads to uncertainties in future projections of SOC stocks and complicates the evaluation of strategies to mitigate climate change through increased SOC sequestration.

In this study we present the parsimonious process-based Carbon Erosion DYNAMics model (CE-DYNAM) that links sediment dynamics resulting from water erosion with the C cycle along a cascade of hillslopes, floodplains and rivers. The model simulates horizontal soil and C transfers triggered by erosion across landscapes and the resulting changes in land-atmosphere CO2 fluxes at a resolution of about 8 km at the catchment scale. CE-DYNAM is the result of the coupling of a previously developed coarse-resolution sediment budget model and the ecosystem C cycle and erosion removal model derived from the ORCHIDEE land surface model. CE-DYNAM is driven by spatially explicit historical land use change, climate forcing, and global atmospheric CO2 concentrations affecting ecosystem productivity, erosion rates and residence times of sediment and C in deposition sites. The main features of CE-DYNAM are (1) the spatially explicit simulation of sediment and C fluxes linking hillslopes and floodplains, (2) the low number of parameters that allow running the model at large spatial scales and over long-time scales, and (3) its compatibility with any global land surface model, hereby, providing opportunities to study the effect of soil erosion under global changes.

We present the model structure, concepts, and evaluation at the scale of the Rhine catchment for the period 1850-2005 AD. Model results are validated against independent estimates of gross and net soil and C erosion rates, and the spatial variability of SOC stocks from high-resolution modeling studies and observational datasets. We show that despite local differences, the resulting soil and C erosion rates, and SOC stocks from our rather coarse-resolution modelling approach are comparable to high-resolution estimates and observations at sub-basin level. The model also shows that SOC storage increases exponentially with basin area for floodplains in contrast to hillslopes as is seen in observations. We find that soil
erosion mobilized 159 Tg ($10^{12}$ g) of C under changing climate and land use, assuming that the erosion loop of the C cycle was in near steady-state by 1850. This caused a net C sink equal to 1% of the Net Primary Productivity of the Rhine catchment over 1850-2005 AD. This sink is a result of the dynamic replacement of C on eroding sites that increases in this period due to rising atmospheric CO$_2$ concentrations enhancing the litter C input to the soil from primary production.

**Keywords:** soil erosion; regional carbon cycle; carbon sink; Rhine catchment

1 Introduction

Soils contain more carbon (C) than the atmosphere and living biomass together. Relatively small disturbances (anthropogenic or natural) to soil C pools over large areas could add up to substantial C emissions (Ciais et al., 2013). With the removal of natural vegetation and the introduction of mechanized agriculture, humans have accelerated soil erosion rates. Over the last two to three decades, studies have shown that water erosion (soil erosion by rainfall and runoff) amplified by human activities has substantially impacted the terrestrial C budget (Doetterl et al., 2012; Lal, 2003; Lugato et al., 2018; Van Oost et al., 2007, 2012; Stallard, 1998; Wang et al., 2017). However, the net effect of water erosion on the C cycle at regional to global scale is still under debate. This leads to uncertainties in the future projections of the soil organic C (SOC) reservoir, and complicates the evaluation of strategies to mitigate climate change by increased SOC sequestration. The study of Stallard (1998) was one of the first to show that water erosion does not only lead to additional C emissions but can also sequester C due to the photosynthetic replacement of SOC at eroding sites and the stabilization of SOC in deeper layers at burial sites. The study of van Oost et al. (2007) was the first to confirm the importance of the sequestration of SOC by agricultural erosion at global scale using isotope tracers. Wang et al. (2017) gathered data on SOC profiles from erosion and deposition sites and confirmed that water erosion on agricultural land that started from the early/middle Holocene has caused a large net global land C sink. Other studies, however, argue that soil erosion is a net C source to the atmosphere due to increased SOC decomposition following soil aggregate breakdown during transport and at deposition sites (Lal et al., 2003; Lugato et al., 2018). Most studies modeling soil erosion and its net effect on SOC dynamics at global scale, however, did not account for the full range of complex effects of climate change, CO$_2$ fertilization increasing productivity and potentially soil C inputs, harvest of biomass, land use change, and changes in cropland management. In addition, models used at large spatial scales mainly focus on hillslopes and removal processes and neglect floodplain sediment and SOC dynamics. This can lead to substantial biases in the assessment of net effects of SOC erosion at catchment scale because floodplains can store substantial amounts of sediment and C (Berhe et al., 2007; Hoffmann et al., 2013a).

Furthermore, soil erosion is one of the main contributors to particulate organic carbon (POC) fluxes in rivers and C export to the coastal ocean. The riverine POC fluxes are usually much smaller than the SOC erosion fluxes, because only a small fraction of eroded material is entering the river network and POC losses in the river network occur due to decomposition and burial on floodplains and in benthic sediments. Therefore, uncertainties in large-scale SOC erosion rates will lead to
even larger uncertainties in lateral C fluxes between land and ocean for past and future scenarios estimated by global empirical models on riverine C export (Ludwig and Probst, 1998; Mayorga et al., 2010).

To address these knowledge gaps, we present a parsimonious process-based modelling approach that integrates sediment dynamics resulting from water erosion with SOC dynamics and the horizontal transport of sediment and C in the continuum from hillslopes, to floodplains and rivers. With this approach we are not only able to simulate lateral soil and C transfers triggered by erosion across landscapes but also the resulting changes in the land-atmosphere CO₂ fluxes. The modelling approach uses a simple sediment budget model which is coupled to SOC erosion removal, C input from litter fall, and SOC decomposition processes diagnosed from the ORCHIDEE global land surface model (LSM) in an offline setting (Naipal et al., 2018). We parameterized and applied the resulting model, known as CE-DYNAM for the Rhine catchment, although it is intended to be made applicable to other large catchments globally. CE-DYNAM combines soil erosion processes, for which small scale differences in topography are of utter importance, with a state-of-the-art representation of large-scale SOC dynamics driven by land use and environmental factors (climate, atmospheric CO₂) as simulated by the ORCHIDEE LSM. The flexible structure of CE-DYNAM makes the model adaptable to the SOC dynamics of any other LSM. In this way it is possible to study the main processes behind the linkages between soil erosion and the global C cycle.

In the next sections we give a detailed overview of CE-DYNAM model structure, the coupling of erosion, deposition and transport with the coarse-resolution SOC dynamics of ORCHIDEE. We then discuss its application for the Rhine catchment, model limitations, uncertainties and its potentials.

2 Methods

2.1 General model description

CE-DYNAM version 1 (v1) is the result of coupling a large-scale erosion and sediment budget model (Naipal et al., 2016) with the SOC scheme of the land surface model ORCHIDEE (Krinner et al., 2005). The most important features of the model are (1) the spatially explicit simulation of lateral sediment and C transport fluxes linking hillslopes and floodplains, (2) consistent simulation of vertical C fluxes coupled with horizontal transport, (3) the low number of parameters that allows running the model at large spatial scales and over long time-scales up to several thousands of years, (4) generic input fields for application to any region or catchment, and (5) compatibility with land surface models (LSMs).

In the ORCHIDEE LSM, terrestrial C is represented by 8 biomass pools, 4 litter pools and 3 SOC pools. Each of the pools varies in space, time and over the 12 Plant Functional Types (PFTs). An extra PFT is used to represent bare soil. Natural and anthropogenic disturbances to the C pools include fire, crop harvest, changes to GPP, litterfall, autotrophic and heterotrophic respiration as a result of climatic changes (Krinner et al., 2005; Guimberteau et al., 2018). The C-cycle processes are represented by a C emulator that reproduces for each PFT all C pools and fluxes between the pools exactly as
in ORCHIDEE in absence of erosion. A net land use change scheme is included in the emulator with mass-conservative bookkeeping of SOC and C input when a PFT is changed into another from anthropogenic land use change (Naipal et al., 2018). The sediment budget model has been added in the emulator to simulate large-scale long-term soil and SOC redistribution by water erosion using coarse-resolution precipitation, land-cover and LAI data from Earth System Models (Naipal et al. 2015, 2016). The C emulator including erosion removal was developed by Naipal et al. (2018) to reproduce SOC vertical profile, removal of soil and SOC starting from the topsoil, and compensatory SOC storage from litter input. As soil erosion is assumed not to change soil and hydraulic parameters but only the SOC dynamics, the emulator allows substituting for the ORCHIDEE model and performing simulations on time scales of millennia with a daily time step, which would be a very computationally expensive or nearly impossible with the full LSM. The concept and all equations of the emulator are described in Naipal et al. (2018). The following subsections describe the different components of the CEDYNA that couples the C and soil removal scheme (Naipal et al., 2018) with the horizontal transport and burial of eroded soil and C (Naipal et al., 2016).

2.2 The soil erosion scheme

The potential gross soil erosion rates are calculated by the Adjusted Revised Universal Soil Loss Equation (Adj. RUSLE) model (Naipal et al., 2015), which is part of the sediment budget model (Fig 1). In the Adj. RUSLE the yearly average soil erosion rate is a product of rainfall erosivity \( R \), slope steepness \( S \), land cover and management \( C_m \) and soil erodibility \( K \):

\[
E = S \times R \times K \times C_m
\]

(1)

The slope-length \( L \) and support practice \( P \) factors, which are part of the original Revised Universal Soil Loss Equation (RUSLE) model (Renard et al., 1997), have been excluded here because their quantification still includes many uncertainties and is not practical for applications at regional to global scales. These factors are a function of local manmade structures and management practices which are difficult to assess for present day and whose changes over the past are even more uncertain. In addition, we focus in this study on potential soil erosion and do not consider erosion-control practices. Naipal et al. (2015) have developed a methodology to derive the slope factor \( S \) and the erosivity factor \( R \) from 5 arcmin resolution data on elevation and precipitation, hereby preserving the high-resolution spatial variability in slope and temporal variability in erosivity. Despite the comparatively coarse resolution of the erosion model, the so derived \( R \) factor was shown to compare well with the corresponding high-resolution product published by Panagos et al. (2017).

2.3 The sediment deposition and transport scheme

The sediment deposition and transport scheme have been adapted from the sediment budget model described by Naipal et al. (2016), which has been calibrated and validated for the Rhine catchment (Fig 1). In the sediment budget model each grid cell contains a floodplain fraction, which is needed to ensure sediment transport between the grid cells (transport from...
one grid cell to another can only follow the connectivity of floodplains). We followed a 2-step methodology to derive floodplains in the Rhine catchment, as soil databases usually do not identify floodplain soil as a separate soil class. First, grid cells were identified that consisted entirely out of floodplains. For this we used the gridded global data set of soil at 5 arcminute resolution, with intact regolith, and sedimentary deposit thicknesses of Pelletier et al. (2016) (Table 1), and identified lowlands and hillslopes based on soil thickness and depth to bedrock. The lowlands were classified as grid cells that contain only floodplains and no hillslopes. Second, we calculated the floodplain fraction of a grid cell ($A_{fl}$) that has both hillslopes and floodplains as a function of stream length and width based on the methodology developed by Hoffmann et al. (2007):

$$A_{fl} = L_{stream} \cdot W_{stream}$$

where $L_{stream}$ is the stream length derived from the HydroSHEDS database (Lehner and Grill, 2013) (Table 1).

$$W_{stream} = a \cdot A_{upstream}$$

Where, $A_{upstream}$ is the upstream catchment area, and $a$ is equal to 60.8, and $b$ is equal to 0.3.

The parameters $a$ and $b$ have been derived from the scaling behavior of floodplain width as estimated from measurements on the Rhine (Hoffmann et al., 2007). The sediment deposition on hillslopes ($D_{hs}$) and floodplains ($D_{fl}$) is calculated as a function of the gross soil removal rates ($E$) with the following equations:

$$D_{fl} = f \cdot E$$

$$D_{hs} = (1 - f) \cdot E$$

$$f = a_f \cdot e^{b_f \cdot \theta_{max}}$$

Where, $f$ is the floodplain deposition factor at 8 km resolution that determines the fraction of gross eroded material transported and deposited in the floodplain fraction of a grid cell. $a_f$ and $b_f$ are constant parameters that relate $f$ to the average topographical slope ($\theta$) of a grid cell depending on the type of land cover. $\theta_{max}$ is the maximum topographical slope of the entire Rhine catchment.

The parameters $a_f$ and $b_f$ are chosen in such a way that $f$ varies between 0.2 and 0.5 for cropland, reflecting the decreased sediment connectivity between hillslopes and floodplains created by manmade structures such as ditches and hedges. For natural vegetation such as forest and natural grassland, $a_f$ and $b_f$ are chosen in a way that $f$ varies between 0.5 and 0.8 assuming that in these landscapes hillslopes and floodplains are well-connected. In each case and within the defined boundaries, the slope gradient determines the final value of $f$. Eroded material that has not been deposited in the floodplains stays on the hillslopes and is assumed to be deposited at the foot of the hillslopes as colluvial sediment.
The floodplain fractions of the grid cells are connected through a 8 km resolution flow routing network (Naipal et al., 2016), where the rivers and streams are indirectly included in the floodplain area but not explicitly simulated. By routing the sediment and C through the floodplain fractions of grid cells we lump together the slow process of riverbank erosion by river dynamics (time scale ≈ a few years to thousands of years), and the rather fast process of transport of eroded material by the rivers (time scale ≈ days). The rate by which sediment and SOC leave the floodplain of a grid cell to go to the floodplain of an adjacent grid cell is determined by the sediment residence time. The sediment residence time (τ) is a function of the upstream contributing area (Flowacc):

\[ \tau = e^{\frac{\text{Flowacc}}{W}} \]  

(6)

The study of Hoffmann et al. (2008) shows that the majority of floodplain sediments have a residence time that ranges between 0 and 2000 years, with a median of 50 years. The constants \( a_i \) and \( b_i \) are chosen in such way that basin \( \tau \) varies between the 5th and 95th percentile of those observations, with a median for the whole catchment of 50 years. These constants are uniform for the whole basin. Floodplain C storage follows the same residence time as sediment on top of the actual decomposition rate of C in a grid cell of ORCHIDEE. The routing of sediment and C between the grid cells follows a multiple-flow routing scheme. In this scheme the flow coming from a certain grid cell is distributed across all lower-lying neighbors based on a weight (\( W \), dimensionless) that is calculated as a function of the contour length (\( c \)):

\[ W_{(i+k,j+l)} = \frac{\theta_{(i+k,j+l)} * c_{(i+k,j+l)}}{\sum_{k,l} |\theta_{(i+k,j+l)} * c_{(i+k,j+l)}|} \]  

(7)

Where \( c \) is 0.5*grid size (m) in the cardinal direction and 0.354* grid size (m) in the diagonal direction. \((i, j)\) is the grid cell in consideration where \( i \) counts grid cells in the latitude direction and \( j \) in the longitude direction. \( i+k \) and \( j+l \) specify the neighboring grid cell where \( k \) and \( l \) can be either -1, 0 or 1. \( \theta \) is calculated as the division between the difference in elevation (\( h \)) give in meters difference and the grid cell size (\( d \)), also in meters:

\[ \theta_{(i+k,j+l)} = \frac{h_{(i,j)} - h_{(i+k,j+l)}}{d} \]  

(8)

The sediment and C routing is done continuously at a daily time-step to preserve numerical stability of the model. More detailed explanation of the methods presented in this section can be found in the study of Naipal et al. (2016).

2.4 Litter dynamics

The four litter pools in the emulator are an below- and an above- ground litter pool, each split into a metabolic and a structural pool with different turnover rates as implemented in ORCHIDEE (Krinner et al., 2005). The belowground litter pools consist mostly out of root residues. Both the biomass and litter pools have a loss flux due to fire as incorporated in ORCHIDEE by the Spitfire model of Thonicke et al. (2010). The litter that is not respired or burnt is transferred to the SOC.
pools based on the Century model (Parton et al., 1987) and the vertical discretization scheme SOC scheme presented by Naipal et al., (2018).

The vertical discretization scheme was introduced in the emulator to account for a declining C input and SOC respiration with depth, and consists of 20 layers with each 10 cm thickness. The litter to soil fluxes from above-ground litter pools are all attributed to the top 10 cm of the soil profile. The litter to soil fluxes from belowground litter pools are distributed exponentially over the whole soil profile according to:

\[ I_{be}(z) = I_{0be} e^{-r \cdot z} \]  

Where \( I_{0be} \) is the below-ground litter input to the surface soil layer and \( r \) is the PFT-specific vertical root-density attenuation coefficient as used in ORCHIDEE. The sum of all layer-dependent litter to soil fractions is equal to the total litter to soil flux as calculated by ORCHIDEE. The vertical SOC profile is modified by erosion and the resulting deposition fluxes, which is discussed in detail the following sections.

2.5 Crop harvest and yield

We adjusted the representation of crop harvest from ORCHIDEE by assuming a variable harvest index for C3 plants that increases during the historical period as shown in the study of Hay (1995) for Wheat and Barley, which are also the main C3 crops in the Rhine catchment. The harvest index is defined by the ratio of harvested grain biomass to above-ground dry matter production (Krinner et al., 2005). In this study the harvest index increases linearly between 0.26 and 0.46 (Naipal et al. 2018) consistent with the average values of Hay (1995). We also found that in certain cases the cropland NPP was too high during the entire period of 1850-2005, especially in the early part of the 20th Century. This is because the cropland photosynthetic rates were adjusted in ORCHIDEE to give a cropland NPP representative of present day values that are higher than for the low input agriculture of the early 20th Century. To derive a more realistic NPP for crop and barley in the Rhine catchment we used the long-term crop yield data obtained from a dataset on 120000 yield observations over the 20th century in Northeast French Départements (NUTS3 administrative division) (Schauberger et al., 2018). According to the yield data assembled by Schauberger et al. (2018), yields in Northeast France for these crops increased fourfold during the last century. Note that crop residues like straw constituted a larger fraction of the total biomass in 1850 than in 2005, but those residues were likely collected and used for animal feed, housing fuel. We did not account for this harvest of residue in the simulation of SOC.

2.6 SOC dynamics without erosion

The change in the carbon content of the PFT-specific SOC pools in the emulator without soil erosion as described by Naipal et al. (2018) (Fig 1):
\[
\frac{dSOC_a(t)}{dt} = \text{litr}_a(t) + k_{pa} \cdot SOC_p(t) + k_{sa} \cdot SOC_s(t) - (k_{ap} + k_{as} + k_{0a}) \cdot SOC_a(t) \tag{10}
\]

\[
\frac{dSOC_p(t)}{dt} = \text{litr}_a(t) + k_{as} \cdot SOC_a(t) - (k_{sa} + k_{sp} + k_{0p}) \cdot SOC_p(t) \tag{11}
\]

\[
\frac{dSOC_s(t)}{dt} = k_{ap} \cdot SOC_a(t) + k_{sp} \cdot SOC_p(t) - (k_{pa} + k_{0p}) \cdot SOC_s(t) \tag{12}
\]

Where, \(SOC_a, SOC_p,\) and \(SOC_s\) (g C m\(^{-2}\)) are the active, slow and passive SOC, respectively. The distinction of these SOC pools, defined by their residence times, are based on the study of Parton et al. (1987). The active SOC pool has the lowest residence time (1 - 5 years) and the passive the highest (200-1500 years). \(\text{litr}_a\) and \(\text{litr}_s\) (g C m\(^{-2}\) day\(^{-1}\)) are the daily litter input rates to the active and slow SOC pools, respectively; \(k_{0a}, k_{0p},\) and \(k_{0s}\) (day\(^{-1}\)) are the respiration rates of the active, slow and passive pools, respectively; \(k_{ap}, k_{sp},\) and \(k_{as}\) are the coefficients determining the flux from the active to the slow pool, from the active to the passive pool, from the passive to the active pool, from the slow to the active pool and from the slow to the passive pool, respectively.

The vertical C discretization scheme in the emulator assumes that the SOC respiration rates decrease exponentially with depth:

\[
k_i(z) = k_{0i} \cdot e^{-re^z} \tag{13}
\]

Where \(k_i\) is the respiration rate at a soil depth \(z\) and \(re\) (m\(^{-1}\)) is a coefficient representing the impact of external factors, such as oxygen availability that decreases with depth. \(k_0\) is the respiration rate of the surface soil layer for a certain SOC pool \(i\).

The variable \(re\) is determined in such a way that the total soil respiration of a certain pool over the entire soil profile without erosion is similar to the output of the full ORCHIDEE model. Detailed description of how this is done can be found in the study of Naipal et al. (2018).

### 2.7 C erosion on hillslopes

In the model we assume that soil erosion takes place on hillslopes, and not in the floodplains due to the usually low topographical slope of floodplains. The factor \((1-f)\) determines the fraction of the eroded soil that is deposited in the colluvial reservoirs (Fig 1). Soil erosion always removes a fraction of the SOC stock in the upper soil layer depending on the erosion rate and bulk density of the soil. The next soil layer contains less C and therefore at the following time-step less C will be eroded under the same erosion rate. To account for this effect, the SOC profile evolution is dynamically tracked in the model and updated at a daily time step, conform with the method of Wang et al. (2015). First, a fraction of the C from each soil pool in proportion to the erosion height is removed from the surface layer. Then, at the same erosion rate, SOC from the subsoil layer becomes the surface layer, maintaining the soil layer thickness in the vertical discretization scheme. Similarly, the SOC from the subsoil later also moves upward one layer. The change in C content due to erosion of the PFT-specific pools for hillslopes can be represented by the following equations:

\[
\frac{dSOC_{HS}(z,t)}{dt} = k_E \cdot SOC_{HS}(z + 1, t) - k_E \cdot SOC_{HS}(z, t) \tag{14}
\]
Where $dSOC_{HS}(z,t) / dt$ is the change in hillslope SOC of a component pool $i$ at a depth $z$ and at time step $t$. The daily erosion fraction $k_E$ (dimensionless) is calculated as following:

$$k_E = \frac{1}{\text{BD} \cdot dz} \cdot \left( \frac{E}{f} \right)$$ \hspace{1cm} (15)

Where, $E$ is the erosion rate (t ha$^{-2}$ year$^{-1}$), $f$ is the floodplain deposition factor, BD is the average bulk density of the soil profile (g cm$^{-3}$) and $dz$ is the soil thickness (=0.1 m).

This part of the model has already applied at the global scale as the C removal model presented by Naipal et al. (2018) and is here extended with the deposition term detailed above.

### 2.8 C deposition and transport in floodplains

The SOC profile dynamics of floodplains are controlled by: (1) C input from the hillslopes, (2) C import by lateral transport from the floodplain fractions of upstream neighboring grid cells, and (3) C export to the floodplain fractions of downstream neighboring grid cells (Fig 1). First, the net eroded flux from the surface layer of the hillslope fraction of the grid cell ($k_E \cdot SOC_{HS}(0)$) is incorporated in the surface layer of the floodplain. At the same deposition rate, the SOC of the surface layer of the floodplain is incorporated in the subsoil layer. Similarly, a fraction of the SOC of the subsoil layer is moved downward one layer. It should be noted that C selectivity is not taken into account here, meaning that the C pools of the deposited material are the same as the eroded material from the topsoil of eroding areas. At the same time a fraction of the C of the surface layer proportional to the sediment residence time ($\tau$) is exported out of the catchment following the sediment routing scheme. This process represents the river bank erosion and resulting POC export by rivers. It should be noted that rivers and streams are not explicitly represented in the model. As we do not have information on the sub-grid spatial distribution of land cover fractions we first sum the exported C flux over all PFTs before assigning the flux proportionally to the land cover fractions of the receiving downstream-lying grid cells. The C that is imported from the neighboring grid cells follows the same procedure as the deposition of eroded material. The change in C content due to deposition and river export/import of the PFT-specific pools for floodplains can be represented by the following equations:

$$\frac{dSOC_{FLi}(z,t)}{dt} = \left( k_D + k_{i\text{out}} \right) \cdot SOC_{FLi}(z-1,t) + \left( \frac{1}{(t+365)} \right) \cdot SOC_{FLi}(z+1,t) - \left( k_D + \frac{1}{(t+365)} + k_{i\text{out}} \right) \cdot SOC_{FLi}(z,t)$$ \hspace{1cm} (16)

for $z>0$

$$\frac{dSOC_{FLi}(0,t)}{dt} = \sum_{n=1}^{n=9} \left( k_{i\text{out}}(n) \cdot SOC_{FLi}(0,t)(n) \right) + \left( k_E \cdot SOC_{HSi}(0,t) \right) + \left( \frac{1}{(t+365)} \right) \cdot SOC_{FLi}(1,t) - \left( k_D + \frac{1}{(t+365)} + k_{i\text{out}}(n) \right) \cdot SOC_{FLi}(0,t)$$ \hspace{1cm} (17)

for $z=0$
Where \( n \) is the neighboring grid cell that flows into the current grid cell, \( d\text{SOC}_{FLi}(z,t) \) is the change in floodplain SOC of a component pool \( i \) at a depth \( z \) and at time step \( t \), and \( \text{SOC}_{HS} \) is the hillslope SOC stock. \( k_D \) is the deposition rate and equal to:

\[
k_D = \frac{k_E \times \text{AREA}_{HS}}{\text{AREA}_{FL}}
\]

(18)

Where \( \text{AREA}_{HS} \) is the hillslope area and \( \text{AREA}_{FL} \) is the floodplain area (m\(^2\)). \( k_{\text{out}} \) is the import rate per C pool \( i \) from neighboring grid cells (dimensionless) and can be calculated as:

\[
k_{\text{out}} = \frac{\sum_{n=9}^{1}(W \times \frac{1}{\text{AREA}_{FL}})(n)}{\text{AREA}_{FL}}
\]

(19)

Where, \( W \) is the weight index of equation 7.

2.9 The land use change bookkeeping model

The land use change bookkeeping scheme includes the yearly changes in forest, grassland and cropland areas in each grid cell as reconstructed by Peng et al. (2017) (Table 1). Peng et al. (2017) derived historical changes in PFT fractions based on LUHv2 land use dataset (Hurtt et al., 2011), historical forest area data from Houghton, and present day forest area from ESA CCI satellite land cover (European Space Agency, ESA, 2014). By using different transition rules and independent forest data to constrain the changes in crop and urban PFTs he derived the most suitable historical PFT maps.

When land use change takes place, the litter and SOC pools of all shrinking PFTs are summed and allocated proportionally to the expanding PFTs, maintaining the mass-balance. In this way the litter pools and SOC stocks get impacted by different input and respiration rates for each soil layer. When forest is reduced, three wood products with decay rates of 1, 10 and 100 years are formed and harvested. The biomass pools of other shrinking land cover types are transformed to litter and allocated to the expanding PFTs. For more details on the land use scheme see the study of Naipal et al. (2018).

2.10 Study-Area

The model is tested for the Rhine catchment (Fig 2), which has a total basin area of 185,000 km\(^2\) covering five different countries in Central Europe. Its large size is beneficial for the application of a coarse-resolution model such as CE-DYNAM to study large-scale regional dynamics in the C cycle due to soil erosion. The Rhine catchment has a very interesting topography, with steep slopes larger than 20\% upstream in the Alps, and large, wide and flat floodplains at the foot of the Alps, the upper Rhine and the lower Rhine. The floodplains store large amounts of sediment and C that originally was eroded from the steep hillslopes upstream. This makes it possible to study the long-term effect of erosion on
hillslope and floodplain dynamics. Furthermore, the Rhine catchment has been experiencing different stages of land use change over the Holocene, with land degradation dating back to more than 5500 years ago (Dotterweich, 2013). In contrast, during the last two decades there has been a general afforestation and soil erosion has been decreasing. These land use changes and changes in erosion make an interesting and important case to study the effect of anthropogenic activities on the C cycle in Europe.

In addition, the Rhine catchment has been the focus of many erosion studies providing observations on erosion and sediment dynamics that can be used for model validation (Asselman, 1999; Asselman et al., 2003; Erkens, 2009; Hoffmann et al., 2007, 2008, 2013a, 2013b; Naipal et al., 2016). The global sediment budget model that forms the basis for the sediment dynamics of CE-DYNAM has been validated and calibrated for the Rhine catchment with observations on sediment storage from Hoffmann et al. (2013b) and derived scaling relationships between sediment storage and basin area (Naipal et al., 2016). These scaling relationships are also applicable for SOC storage and basin area. He found that for floodplains the sediment and C storage increase in a non-linear way with basin area, while hillslopes show linear increase. We will use these relationships to validate the spatial variability in SOC storage of floodplains and hillslopes simulated by CE-DYNAM. The scaling relationships have the form of a simple power law:

\[ M = a \times \left( \frac{A}{A_{ref}} \right)^b \]  

(20)

Where \( M \) is the sediment storage or the SOC storage, \( a \) is the storage (Mt) related to an arbitrary chosen area \( A_{ref} \), while \( b \) is the scaling exponent.

### 2.11 Input data and model simulations

To create the C emulator that forms the underlying C cycle part of CE-DYNAM, we first ran the full ORCHIDEE model for the period 1850-2005 at a coarse resolution of 2.5° degrees latitude and 3.75° degrees longitude, and output all C pools and fluxes. The pools and fluxes were then archived together and used to derive the turnover rates to build the emulator. The SOC scheme of the emulator that has been modified to account for soil erosion processes has been made to run at a spatial resolution of 5 arcmin, similar to the original global sediment budget model. Then, we performed three main simulations with CE-DYNAM for the Rhine catchment. Simulation S0: The baseline simulation or no-erosion simulation, where SOC dynamics are similar to the full ORCHIDEE model. Simulation S1: The erosion-only simulation, where the hillslopes erode and all eroded C is respired to the atmosphere without reaching the colluvial and alluvial deposition sites. Simulation S2: The simulation with full sediment dynamics where hillslopes and floodplains are connected and can bury or loose C. We ran the emulator for 2000 years at a daily time step with the initial climate and land cover of the period 1850-1860. Afterwards, we performed the transient simulations for the period 1851-2005 at a daily time step with changing climate and land cover conditions. However, after 2000 years the model the passive SOC pool was still not in complete equilibrium with a change between 0.8 and 1 g C m\(^{-2}\) year\(^{-1}\). Therefore, we subtracted the additional increase in SOC stocks resulting from the disequilibrium state from the SOC stocks of the transient simulations before analyzing the transient results.
Finally, to ensure a faster performance of CE-DYNAM we delineated the Rhine catchment in 7 large sub-basins based on the flow direction and ran in parallel for each of the sub-basins at the daily timestep. After each year the sub-catchments exchanged the C between each other.

### 3 Results

#### 3.1 Model validation

We performed a detailed model validation of the sediment and the C part of the model based on the following steps: (1) Validation of soil erosion rates using high-resolution model estimates and observations from other studies, (2) validation of C erosion rates using high-resolution estimates for Europe from the study of Lugato et al. (2018), (3) validation of the spatial variability of sediment storage, (4) validation of SOC stocks using data from a global soil database and a European land use survey.

For the validation of gross soil erosion rates we used the high-resolution model estimates from the study of Panagos et al. (2015), who applied the RUSLE2015 model at a 100 m resolution at European scale for the year 2010. The RUSLE2015 is derived from the original RUSLE model with some modifications to the model parameters, especially the L, C and P factors. We also used independent high-resolution erosion estimates from the study of Cerdan et al. (2010), which were based on an extensive database of measured erosion rates under natural rainfall in Europe.

For the validation of net soil erosion rates from hillslopes with those of the study of Panagos et al. (2015), which was extended the RUSLE2015 model with a hillslope sediment deposition and transport scheme based on the sediment delivery ratio concept. We show that the quantile distribution of our gross soil erosion rates fall in between the estimates of these two studies (Fig 3A&B). It should be noticed that our study and the study of Cerdan et al. (2010) simulated potential soil erosion rates, not accounting for erosion control practices that are captured by the P-factor. We also compared the quantile distribution of our net soil erosion rates from hillslopes with those of the study of Lugato et al. (2018), and show that they are similar (Fig 3C). Lugato et al. (2018) extended the RUSLE2015 model with a hillslope sediment deposition and transport scheme based on the sediment delivery ratio concept.

For the comparison of the spatial variability of gross soil erosion rates we used the relationship of erosion to the topographical slope and rainfall erosivity. These two parameters are argued to explain about 70% of the total potential soil erosion rates at regional scales (Doetterl et al., 2012). We show that in our study and the study of Lugato et al. (2018) erosion rates increase with increasing slope and erosivity, and that erosion can be high for very steep slopes with a low erosivity (Fig 4). However, the difference between small and large erosion rates in our study is high, indicating an underestimation of local variability in erosivity and slope. Overall, our results show that our coarse-resolution erosion model is capable of producing reliable estimates of potential soil erosion rates and their spatial variability for the Rhine catchment.
For the validation of gross C erosion rates, we used the results from Lugato et al. (2018), where they coupled the RUSLE2015 erosion model to the Century bio-geochemistry model. They provided an enhanced and a reduced erosion-induced C sink uncertainty scenario, based on different assumptions for C burial and C mineralization during transport. We find that the quantile distribution of our simulated erosion and deposition rates is close to that of Lugato et al. (2018) (Fig 5A-D). We also find that the relation between soil erosion and C erosion rates of our study is similar to the relation of Lugato et al. (2018) and falls within the uncertainty range (Fig 6). The coarse resolution of our model may explain the decreased variability between the estimates.

For the validation of SOC stocks we used the Global Dataset for Earth System Modeling (GSDE) (Shangguan et al., 2014) available at a spatial resolution of 1km and the Land Use/Land Cover Area Frame Survey (LUCAS) (Palmieri et al., 2011). The LUCAS topsoil SOC stocks, available at a high spatial resolution of 500 m, were calculated using the LUCAS SOC content for Europe (de Brogniez et al., 2015) and soil bulk density derived from soil texture datasets (Ballabio et al., 2016). We find that the simulated top 20 cm SOC stocks per land cover type generally fall within the quantile range of the LUCAS SOC stocks (Fig 7). We also find that the simulation with erosion does not substantially change this result but leads to slightly lower SOC stocks due to the impact of erosion and POC export out of the catchment. Furthermore, we find that in both the erosion and no-erosion simulation the SOC stocks for grassland are higher than for forest. This is also observed in the study of Wiesmeier et al. (2012) in South-Germany where they found considerable higher SOC stocks for grassland with a median of 11.8 kg C m$^{-2}$ compared to forest based on the analysis of 1460 soil profiles.

To validate the spatial variability of SOC stocks we delineated 13 sub-basins in the non-Alpine region. We found a realistic spatial variability in topsoil SOC stocks after comparing our simulated SOC stocks from the erosion simulation with the SOC stocks of GSDE and LUCAS at sub-basin level (Table 2).

For the validation of sediment storage in hillslopes and on floodplains we used the same approach as in Naipal et al. (2016), where we based our validation on measured Holocene SOC deposits from the study of Hoffmann et al. (2013). Hoffmann et al. (2013) did an inventory of 41 hillslope and 36 floodplain sediment and SOC deposits related to soil erosion over the last 7500 years. They found that the sediment and SOC deposits were related to the basin size according to certain scaling functions, where floodplain deposits increased in a non-linear way with basin size while the hillslope deposits showed a linear increase with basin size. We selected the grid cells that contained the points of observation of the study of Hoffmann et al. (2013) and found a significantly larger exponent of the scaling relationship between floodplain SOC storage and basin area compared to hillslope SOC storage, which corresponds to the findings of Hoffmann et al. (2013). However, this is not the case when deriving the scaling relationships at sub-basins level (Table 3).

### 3.2 Model application

We find an average annual soil erosion rate of 4.66 t ha$^{-1}$ year$^{-1}$ over the period 1850-2005, which is 1.7 times larger than the average erosion rate simulated for the last millennium (Naipal et al., 2016) and 3.8 times larger than the average...
erosion rate of the Holocene (Hoffmann et al., 2013). This soil erosion flux mobilized around 159 Tg of C over the period 1850-2005, of which 37% is deposited in colluvial reservoirs, 63% is deposited in alluvial reservoirs and 0.2% is exported out of the catchment.

Over the period 1850-2005 there is a general afforestation in the Rhine catchment that started around 1920 AD (Fig 8B). This afforestation takes place in the non-Alpine part of the Rhine and leads to a long-term decreasing trend in gross soil and SOC erosion rates on hillslopes in the non-Alpine region (Fig 8D). In the Alpine part of the Rhine there was a conversion of cropland and forest to grassland. Cropland decreases by 75% over the period 1920 and 1960, while forest decreases by 16% over the period 1910-1950. The conversion of forest to grassland has a stronger impact on the soil erosion rates than the conversion of cropland to forest, resulting in an increase of soil erosion rates over the period 1910-1950 (Fig 8C). This increase is amplified by increased yearly precipitation in this region. Because the soil erosion rates in the Alps are generally much larger than the soil erosion rates in the non-Alpine region due to the steep landscape, the Alpine region dominates the trend in gross soil erosion and C erosion of the entire in the period 1910-1950 (Fig 8C). As a result, the summed gross soil and C erosion rates over the whole Rhine catchment do not show a specific trend (Fig 8C). The temporal variability in the soil and C erosion rates is a result of changes in precipitation (Fig 8A), however, land use change dominates the overall long-term trend. Although precipitation is temporarily very variable, spatially it does not vary significantly over the Rhine.

Furthermore, we find that the temporal variability in C erosion rates follow the soil erosion rates closely, indicating that soil erosion dominates the variations in C erosion over this time-period, while increased SOC stocks due to CO₂ fertilization and afforestation play a secondary role as a slowly, varying trend.

We find that the cumulative C erosion removal flux of 159 Tg leads to a cumulative net C sink of 90 Tg C for the whole Rhine region (Fig 8E). This is about 1% of the cumulative NPP and about one fourth of the cumulative land C sink of the Rhine without erosion. For the non-Alpine part of the Rhine, erosion leads to a net C sink of 55 Tg C, which equals to one fifth of the total land C sink without erosion. It should be noted that these are potential fluxes, assuming that the photosynthetic replacement of C is not affected by the degradation of soil due to the removal of nutrients, declining water-holding capacity and other negative changes to the soil structure and texture (processes not covered by our model). The breaking point in the graphs of figure 8E and F around 1910AD is a result of the climate data used as input.

To better understand the erosion-induced net C flux (Fig 8E, F), we analyzed the erosion-induced C exchange with the atmosphere by creating C budgets for the entire Rhine catchment for the period 1850-1860 and for the period 1950-2005 (Fig 9A&B). These C budgets also shed light on changes in the linkage between lateral and vertical C fluxes over time. As we do not explicitly track the movement of eroded C through all reservoirs (for example between eroding hillslopes and colluvial reservoirs), we make use of the changes in SOC stocks and NEP of the three main simulations (S0, S1, S2) to derive the erosion-induced vertical C fluxes.
By subtracting the Net Ecosystem Productivity of hillslopes (NEP$_{HS}$), which is the difference between NPP and heterotrophic respiration, of the no-erosion simulation (S0) from the erosion-only simulation (S1), we derive the additional photosynthetic replacement of SOC on eroding sites (eq 21):

\[ E_{rep} = NEP_{HS}(S1) - NEP_{HS}(S0) \]  

(21)

Where, \( E_{rep} \) is the potential dynamic Photosynthetic replacement of C on eroding sites (assuming no feedback of erosion on NPP). Part of the eroded C that is transported to and deposited in colluvial reservoirs can be respired or buried (Eq. 22). The difference between NEP of simulation S2 and S1 is the NEP caused by the deposition of eroded C in colluvial areas and equal to the difference between the burial and respiration of C in colluvial sites. As we do not explicitly track the respiration of deposited material in the model, we can only derive the net respiration or net burial of the colluvial deposits (\( R_{cn} \)) with the following equation:

\[ R_{cn} = NEP_{HS}(S2) - NEP_{HS}(S1) \]  

(22)

The same concept can be applied for the net respiration of floodplains:

\[ R_{an} = NEP_{FL}(S2) - NEP_{FL}(S0) \]  

(23)

Where, NEP$_{FL}$ is the floodplain Net Ecosystem Productivity, and \( R_{an} \) is the net respiration or net burial of alluvial deposits. Positive values for \( R_{an} \) or \( R_{cn} \) indicate a net burial or deposition of the deposited material. We find that dynamic replacement of C on eroding sites increased by 39% at the end of the period despite decreasing soil erosion rates (Fig 9A&B). This increase in the photosynthetic replacement of C is due to the globally increasing CO$_2$ concentrations known as the CO$_2$ fertilization effect, amplified by the afforestation trend in the Rhine over this period. Without this fertilization effect, soil erosion and deposition would be likely a weaker C sink or even a C source over the period 1850-2005. Furthermore, we find that the yearly average gross C erosion flux from eroding sites decreased slightly by 2%, while the yearly deposition fluxes in colluvial and alluvial sites decreased by 3.5% and 0.6%, respectively. The decrease in the deposition flux to floodplains is compensated by better sediment connectivity between hillslopes and floodplains due to afforestation. Forests have less man-made structures that can prevent the erosion fluxes from reaching the floodplains, which is represented by a higher floodplain deposition ‘f’ factor in the model.

We also find that colluvial reservoirs show a net respiration flux throughout the time period (Fig 9A&B). This is consistent with previous studies who found that colluvial sites can be areas of increased CO$_2$ emissions (Billings et al., 2019; Van Oost et al., 2012). However, the difference between the start and end of the transient period is triggered by the net respiration/burial flux of deposited C in floodplains. While at the start of the period, deposition in alluvial reservoirs leads to a substantial net burial flux (~0.8 times the floodplain deposition), at the end of the period respiration of deposited SOC in floodplains is larger than this burial flux (Fig 9B). This is a result of an increased respiration of deposited material over...
the entire catchment, most likely due to increasing temperatures over 1850-2005. The constant removal of C-rich topsoil and its deposition in alluvial and colluvial reservoirs makes the deposited sediments generally richer in C than soils on erosion-neutral sites, providing more substrate for respiration. The largest increase in total respiration of alluvial and colluvial deposits takes place in the Alps due to the initial increase in erosion rates resulting in large deposits of C.

We also observe declining erosion rates over the non-Alpine region leading to decreasing or less strong increasing dynamic replacement of C. Both processes, the decrease in burial dynamic C replacement, contribute to a reduced increase in the erosion-induced net C sink over time (Fig 8E, F).

4 Discussion

4.1 Initial conditions and past global changes

Initial climate and land cover/use conditions needed to perform the equilibrium simulation together with the length of the transient period are essential parameters that determine the resulting spatial distribution of soil and C. Landscapes are in a constant transient state due to global changes, such as climate change, land use change, accelerated soil erosion. However, we assumed an equilibrium state so that we can quantify the changes during the transient period. The more one goes back in time to select the initial conditions and the longer the transient period that covers the essential historical environmental changes, the more accurate are the present-day distribution of SOC stocks, sediment storages, and related fluxes. This is especially true when analyzing the redistribution of soil and C as a result of erosion, deposition and transport, as these soil processes can be very slow. For example, the study of Naipal et al. (2016) shows that by simulating the soil erosion processes for the last millennium a spatial distribution of sediment storages that is similar to observations can be found. In this study we modeled steady state initial conditions of the period 1850-1860 due to constraints in data availability on precipitation and temperature, and because the aim of this study is to present the potential and limitations of CE-DYNAM rather than provide precise values for soil and C stocks and fluxes. By focusing only on the period 1850-2005 we miss the effects of significant land use changes in the past that coincided with times of strong precipitation such as in the 14th and 18th century. These major anthropogenic changes in the last Holocene substantially affected the present-day spatial distribution and size of SOC stocks. As a result, we find that floodplains store less SOC than hillslopes at the end of the transient period (Fig 10A), which is different from the findings of Hoffmann et al. (2013) and what can be derived from global soil databases. Hoffmann et al. (2013) showed that the large amount of C stored in the deeper layers of the floodplains can be several thousands of years old. In addition, the high C stocks in floodplains also result from higher local plant productivity due to favorable soil nutrient and hydrological conditions. In our study we do not capture this effect and we do not look at a timescale long enough to capture this distinction between SOC storage in floodplains and in hillslopes. However, we did not find that the vertical distribution of C in floodplain sediment is more homogenous or shows a less strong decrease in C with depth (Fig 10B). This is in line with the findings of previous studies (Hoffmann et al., 2013; Billings et al., 2019).
4.2 Model limitations

Although CE-DYNAM explicitly account for hillslope and floodplains re-deposition, it neglects important processes affecting the C dynamics in floodplains. The model does not account for a slower respiration rate due to low-oxygen conditions, physical and chemical stabilization (Berhe et al., 2008; Martínez-mena et al., 2019) or a higher NPP for nutrient-rich floodplains (Van Oost et al., 2012; Hoffmann et al., 2013). The oxidation and preservation of C in deposition environments, especially in alluvial reservoirs remain highly uncertain (Billings et al., 2019). Furthermore, the model does not take into account the selectivity of erosion, often expressed as the enrichment ratio, where the C content of eroding soil can be higher than that of the original soil. The enrichment ratio can be very variable across landscapes, while the importance of erosion selectivity for C is still under debate (Nadeu et al., 2015; Wang et al., 2010).

CE-DYNAM does not account for different ratios between the SOC pools (active, slow, passive) with depth due to the limitation in information to constrain these fractions for floodplains and hillslopes. However, this can be potentially important for respiration of C in depositional sites and during transport. Studies show that the labile C is decomposed first during sediment transport and directly after deposition, leaving behind the more recalcitrant C in deposition sites (Berhe et al., 2007; Billings et al., 2019). Due to the simplistic nature of our coarse-resolution model and the lack of data on oxidation of eroded C during transport we did not include C respiration during transport in the model. We also do not take into account the transformation of POC to DOC and their fate in rivers and streams. The model also lacks dams in and fixed river banks of the Rhine river. In this way, CE-DYNAM provides only a potential state of soil and SOC redistribution as would be under more natural conditions. Furthermore, there is no feedback between soil erosion and plant productivity in the model. To account for such process soil erosion processes would need to be explicitly included in a land surface model such as ORCHIDEE. The lack of this feedback results in an unlimited dynamic replacement of C on eroding sites.

5 Conclusions

We presented a novel spatially-explicit and process-based C erosion dynamics model, CE-DYNAM, which simulates the redistribution of soil and C over land as a result of water erosion and calculates the role of this redistribution for C budgets at catchment scale. We demonstrate that CE-DYNAM captures the spatial variability in soil erosion, carbon erosion and SOC stocks of the Rhine catchment when compared to high-resolution estimates and observations. We also show that the quantile ranges of erosion and deposition rates and C stocks fall within the uncertainty ranges of previous estimates at basin or sub-basin level. Furthermore, we demonstrate the model ability to disentangle vertical C fluxes resulting from the redistribution of C over land and develop C budgets that can shed light on the role of erosion in the C cycle. The simple structure of CE-DYNAM and the relative low amount of parameters makes it possible to run several simulations to investigate the role of individual processes on the C cycle such as removal by erosion only, or the role of deposition and transport. Its compatibility with land surface models makes it possible to investigate the long-term and large-scale effect of erosion processes under various global changes such as increasing atmospheric CO₂ concentrations, changes to precipitation and temperature, and land use change.
The application of CE-DYNAM for the Rhine catchment for the period 1850-2005 AD reveals three key findings:

- Soil erosion leads to a cumulative net C sink of 90 Tg by the end of the period, which is equal to one fourth of the cumulative land C sink of the Rhine without erosion. This C sink is a result of an increasing dynamic replacement of C on eroding sites due to the CO$_2$ fertilization effect, despite decreasing soil and C erosion rates over the largest part of the catchment. We conclude that it is important to take global changes such as climate change into account to better quantify the net effect of erosion on the C cycle.

- The erosion-induced C sink decreases over time due to decreasing erosion rates and increasing respiration of deposited C in alluvial and colluvial reservoirs. In contrast to colluvial reservoirs, alluvial reservoirs experience a net C burial. However, this net C burial can become net C respiration due to changes in the climate such as global warming. We conclude that burial of eroded C in floodplains plays an essential role in the strength of the erosion-induced C sink.

- Initial climate and land cover conditions and the transient period over which erosion under global changes takes place are essential for the determination if soil erosion is a net C sink or source and to what extent.

Altogether, these results indicate that despite model uncertainties related to the relative coarse spatial resolution, missing or simplified processes, CE-DYNAM represents an important step forwards into integrating soil erosion processes and sediment dynamics in Earth system models.

**Code and data availability**

The source code of CE-DYNAM is included as a supplement to this paper. Model data can be accessed from the Zenodo repository under the doi:10.5281/zenodo.2642452 (not published yet). For the other data sets that are listed in Table 1, it is encouraged to contact the first authors of the original references.

**Author contributions**

VN built and implemented the model. YW provided the basic structure for the model. All authors contributed in the interpretation of the results and wrote the paper.

**Competing interests**

*The authors declare that they have no conflict of interest.*

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Table 1: Model input datasets

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<th>Temporal resolution</th>
<th>Period</th>
<th>Source</th>
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<tr>
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<td>6 hourly</td>
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<tr>
<td>precipitation for the Adj. RUSLE</td>
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<td>monthly</td>
<td>1850-2005</td>
<td>ISIMIP2b (Frieler et al., 2017)</td>
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<td>Soil</td>
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<td>-</td>
<td>Global Soil Dataset for Earth System Modeling, GSDE (Shangguan H.W., Dai Y., Duan Q., Liu B., 2014)</td>
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<td>Topography</td>
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<td>-</td>
<td>GTOPO30; U.S. Geological Survey, EROS Data Center Distributed Active Archive Center 2004; <a href="https://www.ngdc.noaa.gov/mgg/topo/gltile.shtml">https://www.ngdc.noaa.gov/mgg/topo/gltile.shtml</a>; last access: 5 April 2019</td>
</tr>
<tr>
<td>Flow accumulation</td>
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<td>-</td>
<td>HydroSHEDS (Lehner et al., 2013); <a href="https://www.hydrosheds.org/">https://www.hydrosheds.org/</a>; last access: 5 April 2019</td>
</tr>
<tr>
<td>Hillslopes/Floodplain area</td>
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<td>-</td>
<td>Pelletier et al. (2016)</td>
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<td>River network &amp; stream length</td>
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<td>-</td>
<td>Hydrosheds (Lehner et al., 2008)</td>
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Table 2: This table shows the results of the linear regression between the simulated total SOC stocks (Tg of C per year) and those of the Global Soil dataset for Earth System Modeling (GSDE) and from the LUCAS database. The regression is done after aggregating the data at sub-basin level for the 13 sub-basins that were delineated in the Rhine catchment. RMSE is the root mean square error given in Tg of C per year, while the r-value is the spatial correlation coefficient.

<table>
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<th>p-value</th>
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<tr>
<td>This study versus GSDE</td>
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<td>&lt;0.01</td>
<td>29.32</td>
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</table>

Table 3: This table presents the scaling exponent (b) of equation 20 for floodplains and hillslopes. The scaling exponent was derived for selected points in the Rhine catchment for which measurements on the SOC storage were taken by Hoffmann et al. (2013), and at sub-basin level after the data on area and SOC stocks was aggregated for each of the 13 sub-basins of the Rhine.

<table>
<thead>
<tr>
<th></th>
<th>Scaling exponent floodplains</th>
<th>Scaling exponent hillslopes</th>
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<td>Hoffmann et al. (2013)</td>
<td>1.23±0.06</td>
<td>1.08±0.07</td>
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Figure 1: A conceptual diagram of CE-DYNAM. The red arrows represent the C fluxes between the C pools/reservoirs, while the black arrows represent the link between the erosion processes (removal, deposition and transport).
Figure 2: The Rhine catchment (Hoffmann et al., 2013)

Figure 3: Quantile-whisker plot of simulated gross soil erosion rates (t/year) (grey whisker boxes), compared to (A) the study of Cerdan et al. (2010) and (B) the study of Panagos et al. (2015) (orange whisker boxes). (C) Quantile-whisker plot of simulated net soil erosion rates (t/year) (grey whisker boxes), compared to the study of Borrelli et al. (2018) (orange whisker boxes). Medians are plotted as red horizontal lines.
Figure 4: Normalized gross soil erosion as a function of the topographical slope (%) and rainfall erosivity for (A) this study and (B) RUSLE2015.
Figure 5: (A) Hillslope C erosion rates and, (B) C deposition rates, compared to the enhanced erosion scenario from Lugato et al. (2018). (C) Hillslope C erosion rates and, (D) C deposition rates, compared to the reduced erosion scenario from Lugato et al. (2018).

Figure 6: The relationship between soil erosion and C erosion of simulation S2 (black points) in comparison to the erosion scenarios from the study of Lugato et al. (2018) with enhanced (red points) and reduced erosion (blue points), respectively. The straight lines are the trendlines of the linear regression between soil and C erosion.
Figure 7: Comparison of total SOC stocks per land cover type between the simulation without erosion (red boxes), the simulation with erosion (black boxes) and the LUCAS data (green boxes). The red horizontal lines are the medians, the blue stars are the means, the dashed vertical lines represent the range between the minimum and maximum, and the black dots are the outliers.

Figure 8: Timeseries of (A) the 5-year average yearly precipitation of the entire Rhine catchment (mm), (B) changing land cover fractions of the entire Rhine catchment, (C): 5-year average of the total gross soil erosion (Pg year\(^{-1}\)) and C erosion rates (Tg C year\(^{-1}\)) of the entire Rhine catchment, (D): 5-year average of the total gross soil erosion (Pg year\(^{-1}\)) and C erosion rates (Tg C year\(^{-1}\)) of the non-Alpine region of the Rhine. Erosion on bare soil is not taken into account here. (E) Cumulative C emissions from the soil to the atmosphere under land use change and climate change without soil erosion (\(F_{\text{atm0}}\)), with soil erosion (\(F_{\text{atm1}}\)), due to additional respiration or stabilization of buried soil and photosynthetic replacement of C under erosion (Ep) of the entire Rhine catchment. (F) Cumulative C emissions from the soil for the non-Alpine part of the Rhine catchment. Positive values indicate net C emissions to the atmosphere and negative values indicate net C uptake from the atmosphere by the soil.
Figure 9: (A) C budget of the entire Rhine for the period 1851-1861, and (B) for the period 1995-2005. The budget shows the net exchange of C (Tg C year\(^{-1}\)) between the soil and atmosphere as a result of accelerated soil erosion rates. Grey arrows are the erosion-induced yearly average vertical C fluxes, while the brown arrows are the erosion-induced yearly
average lateral C fluxes. The grey boxes represent yearly average changes in SOC stocks for the specific time period as a result of land use change, climate change, erosion and deposition. Dc: Deposition of C on hillslopes; Da: Deposition of C in floodplains; POC\text{exp}: net POC export flux; NEP\text{HS}: Net ecosystem productivity of hillslopes; NEP\text{FL}: Net ecosystem productivity of floodplains; dSOC: Yearly average change in the total SOC stock; dSOC\text{HS}: Yearly average change in the hillslope SOC stock; dSOC\text{FL}: Yearly average change in the floodplain SOC stock.

\textbf{Figure 10:} (A) Vertical distribution of hillslope (red) and floodplain (blue) SOC stocks (kg m\(^{-2}\)) with depth averaged over the whole Rhine catchment, and (B) the vertical distribution of normalized hillslope (red) and floodplain (blue) SOC stocks (dimensionless) with depth averaged over the whole Rhine catchment.