Evaluation of empirical approaches to estimate the variability of erosive inputs in river catchments

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Zusammenfassung

Die Dissertation erforscht die Unsicherheit, Sensitivität und Grenzen großskaliger Erosionsmodelle. Die Modellierung basiert auf der allgemeinen Bodenabtragsgleichung (ABAG), Sedimenteintragsverhältnissen (SDR) und europäischen Daten. Für mehrere Regionen Europas wird die Bedeutung der Unsicherheit topographischer Modellparameter, ABAG-Faktoren und kritischer Schwebstofffrachten für die Anwendbarkeit empirischer Modelle zur Beschreibung von Sedimentfrachten und SDR von Flusseinzugsgebieten systematisch untersucht.

Der Vergleich alternativer Modellparameter sowie Kalibrierungs- und Validierungsdaten zeigt, dass schon grundlegende Entscheidungen in der Modellierung mit großen Unsicherheiten behaftet sind. Zur Vermeidung falscher Modellvorhersagen sind daher kalibrierte Modelle genau zu dokumentieren. Auch wenn die geschickte Wahl nicht-topographischer Algorithmen die Modellgüte regionaler Anwendungen verbessern kann, so gibt es nicht die generell beste Lösung.

Die Auswertungen zeigen auch, dass SDR-Modelle stets mit Sedimentfrachten und SDR kalibriert und evaluiert werden sollten. Mit diesem Ansatz werden eine neue europäische Bodenabtragskarte und ein verbessertes SDR-Modell für Flusseinzugsgebiete nördlich der Alpen und in Südosteuropa abgeleitet. Für andere Regionen Europas sind die SDR-Modelle limitiert. Die Studien zur jährlichen Variabilität der Bodenerosion zeigen, dass jahreszeitlich gewichtete Niederschlagsdaten geeigneter als ungewichtete sind.

Trotz zufriedenstellender Modellergebnisse überwinden weder sorgfältige Algorithmenwahl noch Modellverbesserungen die Grenzen europaweiter SDR-Modelle. Diese bestehen aus der Diskrepanz zwischen den modellierten Bodenabtrags- und den maßgeblich zur beobachteten bzw. kritischen Sedimentfracht beitragenden Prozessen sowie der außergewöhnlich hohen Sedimentmobilisierung durch Hochwässer. Die Integration von nicht von der ABAG beschriebenen Prozessen und von Starkregentagen sowie die Disaggregation kritischer Frachten sollte daher weiter erforscht werden.

Schlagwörter:

allgemeine Bodenabtragsgleichung (ABAG), Modellsensitivität, Modellunsicherheit, Sedimenteintragsverhältnis (SDR), Sedimentfracht

Abstract

This dissertation thesis addresses the uncertainty, sensitivity and limitations of large-scale erosion models. The modelling framework consists of the universal soil loss equation (USLE), sediment delivery ratios (SDR) and European data. For several European regions, the relevance of the uncertainty in topographic model parameters, USLE factors and critical yields of suspended solids for the applicability of empirical models to predict sediment yields and SDR of river catchments is systematically evaluated.

The comparison of alternative model parameters as well as calibration and validation data shows that even basic modelling decisions are associated with great uncertainties. Consequently, calibrated models have to be well-documented to avoid misapplication. Although careful choices of non-topographic algorithms can also be helpful to improve the model quality in regional applications, there is no definitive universal solution.

Further analyses also show that SDR models should always be calibrated and evaluated against sediment yields and SDR. With this approach, a new European soil loss map and an improved SDR model for river catchments north of the Alps and in Southeast Europe are derived. For other parts of Europe, the SDR models are limited. The studies on the annual variability of soil erosion reveal that seasonally weighted rainfall data is more appropriate than unweighted data.

Despite satisfactory model results, neither the careful algorithm choice nor model improvements overcome the limitations of pan-European SDR models. These limitations are related to the mismatch of modelled soil loss processes and the relevant processes contributing to the observed or critical sediment load as well as the extraordinary sediment mobilisation during floods. Therefore, further research on integrating non-USLE processes and heavy-rainfall data as well as on disaggregating critical yields is needed.

Keywords:

model sensitivity, model uncertainty, sediment delivery ratio (SDR), sediment yield, universal soil loss equation (USLE)

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Abbreviations and variables

A	Catchment area, in km^2
A	Interpolation method for suspended-solids data (time-weighted SSC), as index in chapter 3
A_{500}	Catchments in region North with A<500 km^2 (region defined in chapter 5)
a	Coefficients of regression model $y=a_2x^2 + a_1x$
a	year (unit)
Ar	Fraction of arable land in catchments, in %
ATKIS	Amtliches Topographisch-Kartographisches Informationssystem (Official Topographic-Cartographic Information System in Germany)
B	Interpolation method for suspended-solids data (time-weighted SS loads), as index in chapter 3
B	Retention term (chapter 4), in $Mg \cdot km^{-2} \cdot a^{-1}$
C	Cover and management factor of the USLE, -
CCM2	Catchment characterisation and modelling (dataset) v2.0
CLC	Corine Land Cover (dataset)
CSE	Central-South East Europe (region defined in chapter 5, Tab. 28)
D8	Flow-routing algorithm to model water flow in raster DEM (8 cardinal directions)
D∞	Flow-routing algorithm to model water flow in raster DEM (infinite directions)
DD	Drainage density in a catchment, DD = flow length / A, in $km \cdot km^{-2}$
DE	Germany (region defined in chapter 5)
DEM	Digital elevation model
E	Soil loss, E = SDR · SY, in $Mg \cdot km^{-2} \cdot a^{-1}$
FSD	Functional streamflow disaggregation approach to disaggregate water discharge Q (chapter 3)
GIS	Geographical information system

GPCC	Global Precipitation Climatology Centre (source of a pan-European rainfall dataset)
H	Height or average height in catchment, in m
HI	Hypsometric integral of a catchment, HI = $(H - H_{min})/(H_{max} - H_{min})$, -
HISTALP	Historical instrumental climatological surface time series of the greater Alpine region
H_{max}	Maximum height of catchment, in m
H_{min}	Minimum height of catchment, in m
K	Soil erodibility factor of the USLE, in $Mg{\cdot}h{\cdot}ha{\cdot}N^{-1}$ or $Mg{\cdot}ha{\cdot}h{\cdot}ha^{-1}{\cdot}MJ^{-1}{\cdot}mm^{-1}$
L	Slope-length factor of the USLE, -
L·S	Topographic factor of the USLE, -
L_{100m}	L factor derived with erosive slope length of 100 m (chapters 3 and 5), -
L_{emp}	L factor derived with an empirical model (based on measured values) (chapters 3 and 5), -
L_{GIS}	L factor iteratively derived from DEM (chapters 2 and 3), -
LRC	Lech River catchment (chapter 4)
LVC	Land vegetation cover (chapter 4)
MAE	Mean absolute error
ME	Nash-Sutcliffe model efficiency
MFI	Modified Fournier Index, in $mm{\cdot}a^{-1}$
n	Sample size
NRW	(German Federal State of) Nordrhein-Westfalen (North-Rhine Westphalia)
NUTS	Nomenclature des unités territoriales statistiques (Nomenclature of territorial units for statistics)
P	Support practice factor of the USLE, -
PESERA	Pan-European soil erosion risk assessment (soil-loss model and soil-loss map)
Pr	Precipitation, (average) annual precipitation (chapters 3–5), in mm or $mm{\cdot}a^{-1}$

Pr_M	Monthly precipitation (chapters 3 and 4), in mm
Q	Water discharge, in $m^3 \cdot s^{-1}$
q	Area-specific water discharge, q = Q / A, in $mm \cdot a^{-1}$ or $l \cdot s^{-1} \cdot km^{-2}$
Q_{crit}	Critical Q, threshold above which average SSC and SY increase, used to calculate SY_{graph} (chapter 3)
Q_{fast}	"Fast" component of Q as obtained by the FSD approach, $m^3 \cdot s^{-1}$
R	Rainfall and runoff factor of the USLE, in $N \cdot h^{-1} \cdot a^{-1}$ or $MJ \cdot mm \cdot ha^{-1} \cdot h^{-1} \cdot a^{-1}$
r	Pearson product-moment correlation coefficient, -
r_S	Spearman's rank correlation coefficient (Spearman's rho)
RAMSES	Rainfall model for sediment yield simulation (chapter 4)
S	Slope-steepness factor of the USLE, -
SCA	Specific catchment area, SCA = A / unit contour width (or raster-cell width), in $m^2 \cdot m^{-1}$
SDR	Sediment delivery ratio, SDR = E / SY, relative or in %
SE	South-east Europe (region defined in chapter 5, Tab. 28)
SRTM	Shuttle radar topography mission (a DEM source)
SS	Suspended solids
SSC	Suspended-solids concentration, in $kg \cdot m^{-3}$
STI	Sediment transport capacity index, $STI = (SCA/22.13)^{0.6} \cdot (\sin \beta/0.0896)^{1.3}$, -
SW	South-west Europe (region defined in chapter 5, Tab. 28)
SY	Sediment or suspended-solids yield, SY = SSC · q, in $Mg \cdot km^{-2} \cdot a^{-1}$
SY_{alt_bin}	Alternative SY_{graph} (in chapter 3), in $Mg \cdot km^{-2} \cdot a^{-1}$
SY_{ann_graph}	Alternative SY_{graph} (in chapter 3), in $Mg \cdot km^{-2} \cdot a^{-1}$
SY_{base}	Alternative SY_{graph} (in chapter 3), in $Mg \cdot km^{-2} \cdot a^{-1}$
SY_{graph}	Critical SY values based on the graphical-statistical approach (in chapter 3), in $Mg \cdot km^{-2} \cdot a^{-1}$

SY_{FSD}	Critical SY obtained with the FSD approach, $SY_{FSD} = q_{fast} \cdot SSC$ (in chapter 3), in $Mg{\cdot}km^{-2}{\cdot}a^{-1}$
SY_{mod}	Modelled SY, in $Mg{\cdot}km^{-2}{\cdot}a^{-1}$
tot	Total (SY or Q), as index
USLE	Universal soil loss equation, $E = R \cdot K \cdot C \cdot P \cdot L \cdot S$
W	West Europe (region defined in chapter 5)
β	Slope angle, in ° or %
β_{max}	β calculated from the maximum elevation change in the neighbourhood of a DEM raster cell, ° or %
β_{Nbh}	β calculated from a plane fitted to the neighbourhood of a DEM raster cell, ° or %
σ	Standard deviation
τ	Kendall's tau coefficient (Kendall's rank correlation coefficient)

Note: The catchment parameters used for correlation analyses in chapter 3 are listed in Tab. 13 (p. 59). The acronyms for the alternative USLE soil loss maps evaluated in chapter 5 are explained in Tab. 21 (p. 91).

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1 Introduction

Examples of soil erosion. Left: sheet erosion and relocation of topsoil after rainfall on a barren, gentle slope in the German Federal State of Brandenburg, right: high sediment load in the Alpine upper Lech River (Austria) after a thunderstorm

1.1 Soil erosion

1.1.1 The relevance of soil erosion

Soil erosion and sediment fluxes influence the pedo- and hydrosphere (Gr. *pedon* = soil), land- and riverscapes (Allan 2004) as well as bio- and geo-ecosystems (Tab. 1). Soil erosion (Lat. *erodere* = to gnaw off) is a natural process controlled by climate, vegetation and other exogenous factors that has been shaping landscapes over geological time scales. During the last millennia, however, human activity – namely agriculture and deforestation – has increased global soil erosion rates far beyond natural soil formation rates (Wilkinson and McElroy 2007; Dotterweich 2008; Dreibrodt et al. 2010; Giguet-Covex et al. 2011).

This accelerated and enhanced soil erosion has altered global nutrient and carbon cycles (Filippelli 2008; Kuhn 2010; Quinton et al. 2010), led to degraded soils (Millennium Ecosystem Assessment 2005; Jones et al. 2012) as well as lower soil quality and productivity (Banwart 2011). Due to the latter, soil erosion has posed a serious threat to crop production and food supply in the past, as well as today and in the future, in large parts of world (Oldeman 1992; EEA 2006; Pimentel 2006; Bakker et al. 2007; Montgomery 2007; Wilkinson and McElroy 2007; Dotterweich 2008; Ye and van Ranst 2009)

Excessive sedimentation of eroded soil particles in surface water can degrade habitats by clogging gravel beds and reducing the light availability. The stress exerted on flora and fauna depends on the amount and composition of the eroded soil as well as the magnitude, timing and frequency of erosion events (Bilotta and Brazier 2008). High sediment loads in rivers also affect the morphology of rivers and lakes, navigation, fishery and flood hazard (Owens 2005). Sediment deficit in rivers due to sediment trapping and gravel mining, on the other hand, can enforce severe channel incisions, thus undermining bridges and other infrastructure, and diminishes the sediment supply for river deltas and coastlines (Taylor et al. 2008; Batalla and Vericat 2010).

Soil erosion and degradation have resulted in decades of intensive scientific exploration (Schmidt et al. 2010). The need to protect the functions of soil and water, to promote standards for good agriculture and ecological conditions, and to mitigate human impact on soil and water resources has led to the many management options which exist today (Powlson et al. 2011). Nonetheless, these authors also clearly state that "…it is the adoption of erosion control methods rather than their availability that is lacking…" (p. S81). In the European Union, declarations, guidelines, legal actions, and economic incentives are established at different levels to promote sustainable soil use and to protect water bodies (Fullen 2003; Creamer et al. 2010) (Tab. 2). Monitoring, sediment budgets, risk assessment, and modelling are at hand to evaluate how efficient counter-measures are – all methods differing in scale applicability, data requirement and complexity (Cherry et al. 2008).

Tab. 1: Exemplary relationships between soil erosion and society, terrestrial and aquatic bio-geo-systems

System	Relationship	Reference
Soil	Soils less fertile and ploughable in small Central European catchments, eroded top soils and buried top soils downslope	Dotterweich (2008)
	Holocene floodplain sedimentation rates in the Rhine catchment strongly affected by human activity for 3 millennia	Hoffmann et al. (2009)
	Erosion affects both carbon emission and sequestration	Quinton et al. (2010)
	Direct and indirect consequences of soil erosion on soil biodiversity, soil biodiversity influences erodibility	Jeffery et al. (2010), Pohl et al. (2009)
Aquatic habitat and life	Trophic status of water bodies depends on composition and amount of eroded soil	Ekholm and Lehtoranta (2012)
	Abundance and distribution of mussels affected by quantity and the composition of sediments	Brim Box and Mossa (1999)
	Microbial activity degrading coral reefs triggered by organic matter in terrestrial sediments	Weber et al. (2012)
	Intensity, duration and frequency of sediment exposure related to impact on corals, large range of tolerance of coral reefs	Erftemeijer et al. (2012)
	Declined trout reproduction possibly related to sediment load in lowland and alpine rivers	Scheurer et al. (2009)
	Walleye eggs relatively tolerant to the exposure of suspended sediment	Suedel et al. (2012)
	Degraded fish communities related to suspended sediment and agriculture	Meador and Goldstein (2003)
Economy and agriculture	About 80% of global agricultural land affected by moderate to severe soil erosion	Pimentel (2006)
	Annual damage of soil erosion in the U.S.A. above $44 \cdot 10^9$ USD, costs of $38 \cdot 10^9$ EUR for the EU (25 countries) in 2003	Pimentel et al. (1995), Montanarella (2007)
	Soil erosion increases flood damage	Hilker et al. (2009), (Merz et al. 2010)
	Benefit of local communities and aquaculture from the shoreline protection as one of the mangrove ecosystem services estimated to be $1.9 \cdot 10^6$ IDR per household in 1991 for Bintuni Bay, West Papua, Indonesia	Ruitenbeek (1992), Vo et al. (2012)
	Accelerated sediment accumulation in European lakes since mid-20th century, 0.5–1% annual loss of global storage capacity of dams	Rose et al. (2011), WCD (2000)

Tab. 2: Legislative acts in Europe related to soil loss and sediment in surface waters

Target	Region	Legislative act	Made legal	Exemplary focus, aim or regulation
Soil	Austria	Soil protection acts in 5 federal states	1987–2001	Soil loss mitigation, monitoring
	Germany	Federal Soil Protection Act, Soil Protection and Contaminated Sites Ordinance	1999	Soil loss mitigation, monitoring
	Netherlands	Soil Protection Act	1987	Regulations to avoid damage due to soil loss
	Switzerland	Ordinance on Soil Protection	1998	Crop rotation, soil fertility
	Alpine states	Alpine Convention and soil conservation protocol	1995	Soil conservation, erosion control
	Rhine riparian states	Convention on the Protection of the Rhine	2003	Diffuse nutrient sources, flow management (considering natural flow of solid matter)
	European Union	Thematic Strategy for Soil Protection	2006	Sustainable use of soil, soil functions, soil degradation, restoration and protection
Water	European Union	Freshwater Fisheries Directive	2006	Average concentration of suspended solids in salmonid and cyprinid waters ≤25 mg l^{-1}
		Water Framework Directive	2000	Good ecological potential and surface water chemical status, quality elements (river continuity, morphology, biology), suspended material as main pollutant
Environment (general)	Switzerland	Direct payment scheme	1993, new since 1998	Incentives for ecological services, ecological performance (e.g., soil conservation)
	European Union	Common Agricultural Policy (cross-compliance)	2003, new since 2009	Direct support if farmers maintain good agricultural and environmental conditions
		European Agricultural Fund for Rural Development	2005	Financial support for agri-environmental commitments

Owens (2005) argues that sediment is best managed at the catchment scale because local activities like dam construction affect areas and users downstream and because important diffuse sources are dispersed over large areas. Models are especially useful on such a large scale, yet Cherry et al. (2008) recommend combining the advantages of different assessment methods.

1.1.2 Processes, scales, and variability

Soil erosion refers to the mobilisation of soil particles and their transport as sediment by some agent (water, wind) across the landscape including deposition and remobilisation. It is distinguished from mass movements of soil under gravitational influence like landslides. In this dissertation thesis, the topic is soil erosion by water as the most important soil erosion type in Europe and worldwide (Oldeman 1992; Quinton et al. 2010). Oldeman (1992) estimated that an area of $114 \cdot 10^6$ ha in Europe and $1.1 \cdot 10^9$ ha globally is affected. Jones et al. (2012) recently deducted a similar area of $130 \cdot 10^6$ ha for the EU-27. Quinton et al. (2010) obtained a global sediment flux of $35 \cdot 10^9$ $Mg \cdot a^{-1}$ which equals to 5 $Mg \cdot a^{-1}$ for every single person, and assigned 80% to soil erosion by water. Apart from the dominating water, forces like wind, gravity, tillage, and crop harvesting are also of local and regional relevance for soil loss in catchments and sediments in rivers (Boardman and Poesen 2006b; Jones et al. 2012).

Soil erosion by water starts with the detachment and entrainment of soil particles due to the kinetic energy of raindrops and the shear force of runoff. While splash erosion is only effective over short distances, overland flow (either due to snow-melt, infiltration or saturation excess) can transport soil particles much further. Detachment, deposition and remobilisation vary in space and time according to flow (volume, velocity, turbulence), surface, soil and sediment characteristics such as grain size and shear strength. Soil erosion can occur either uniformly as sheet erosion (rain splash, shallow overland flow) or linearly as rill and gully erosion (preferential overland flow, tunnel erosion).

The many geological, biological, anthropogenic, and hydro-climatic processes and factors involved in soil erosion operate at very different scales (Lane et al. 1997; Habersack 2000; Renschler and Harbor 2002; Hinderer 2012). Complex scale relationships (de Vente and Poesen 2005; de Vente et al. 2007) impede the simple transfer of erosion rates and controlling factors across scales and imply the need to carefully consider scale when measuring and modelling soil erosion (Govers 2011; Hinderer 2012). For example, erosion rates estimated from riverine sediment yields are usually far below gross soil loss within the catchment (Walling 1983; Walling and Webb 1996; Wilkinson and McElroy 2007) and the relationships between sediment export and soil loss are loose in space and time (Walling 1983). Similarly, short-term soil erosion rates can considerably differ from long-term values (Parsons et al. 2004; Parsons et al. 2006; Peeters et al. 2008). The influence of opposing human activity like crop production and mining as well as dam construction and soil conservation on decadal trends in sediment yields of the world’s rivers is shown in Walling (2008).

In this dissertation thesis, soil erosion is considered at the catchment scale, i.e. the hydrological catchments of rivers, lakes and reservoirs. Commonly, inverse relationships between area and sediment discharge per unit area and time, in other words the sediment yield or net erosion, have been observed (as conceptualized by de Vente and Poesen (2005) and reviewed by de Vente et al. (2007)). Nonetheless, this concept has been challenged by contrary findings in densely vegetated, sparsely cultivated catchments and river valleys with easily erodible sediments (Walling and Webb 1996; Dedkov 2004). Lu et al. (2005) also revealed that the spatial pattern of erosion sources can control the scale dependency of sediment yield.

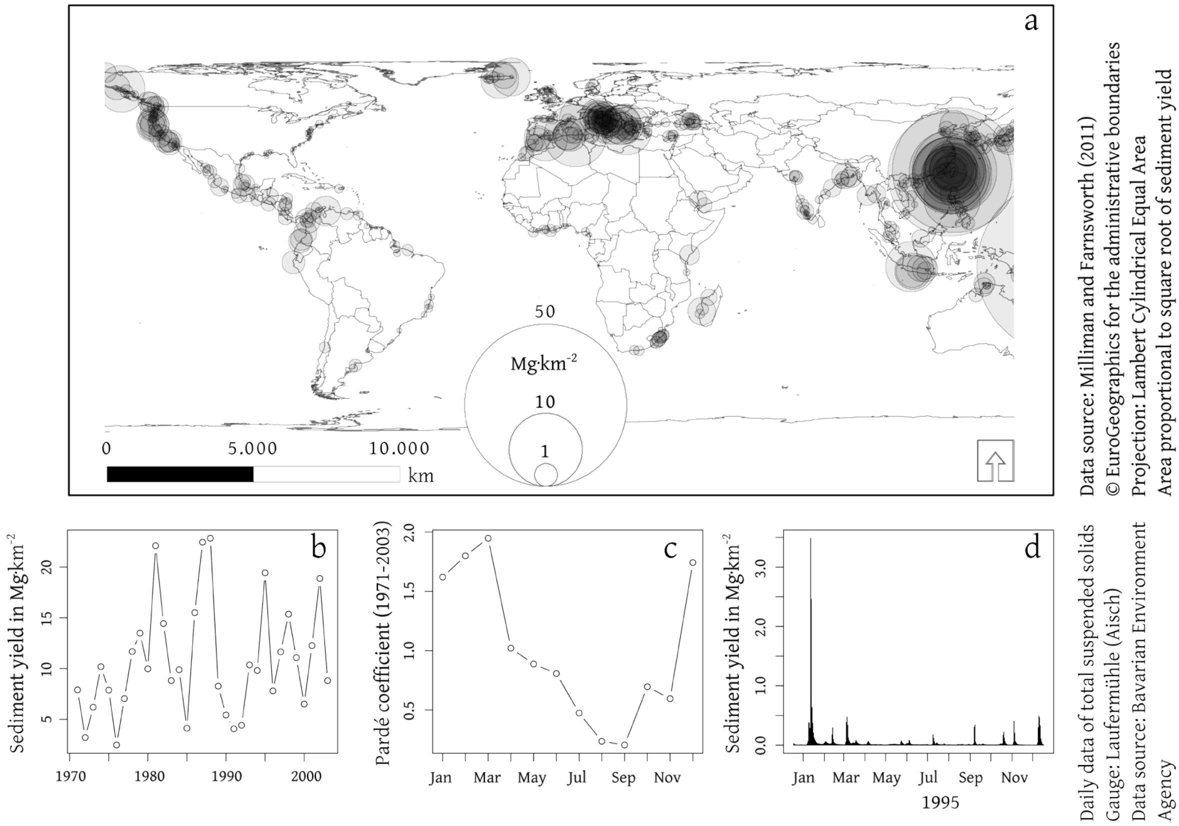

Fig. 1: Spatial and temporal variability of net erosion of river catchments (sediment yield) at different scales. a) average annual yield of large river catchments, b-d) annual, monthly (as ratio of average monthly and annual sum), and daily yield

The common decrease of sediment yields in larger areas is explained with gentler slopes and floodplains in downstream areas (Walling and Webb 1996; Hinderer 2012). Sediment is stored as colluvium at the foot of hillslopes, as alluvial fans in valleys, and in floodplains of rivers. According to Wilkinson and McElroy (2007), sediment accumulation accounts for the striking discrepancy of the global sediment flux from croplands ($75 \cdot 10^9$ Mg·a^{-1}) and the flux derived from sediment load and rock volume ($21 \cdot 10^9$ Mg·a^{-1}). For instance, the Upper Rhine Valley is a short- and long-term sediment sink and de-couples the Upper and High Rhine from the Middle and Lower Rhine (Hoffmann et al. 2007). Many studies have confirmed that sinks can dominate sediment budgets of catchments at different spatial and temporal scales (Trimble and Crosson 2000; Walling and Owens 2003; Walling et al. 2011; Houben 2012).

The numerous processes and factors involved make soil erosion rates highly variable in space and time (Fig. 1). For instance, global and continental sediment yields of river catchments vary over orders of magnitude (Walling and Webb 1996; Milliman and Farnsworth 2011; Vanmaercke et al. 2011) depending on environmental conditions and human activity. Within river catchments, a few sites (the critical source areas) can contribute disproportionately to sediment yields (Heathwaite et al. 2000; Heckrath et al. 2008; Kovacs et al. 2012). The distribution of erosion rates during single events and years is also skewed (Morgan 2005) and a few events can dominate (multi-)annual and even long-term soil erosion rates (Lamoureux 2002; González-Hidalgo et al. 2009; González-Hidalgo et al. 2010; Duvert et al. 2011; Prasuhn 2011). Due to the complexity of soil erosion and the high variability, monitoring of soil erosion is complex and costly and, thus, often complemented by modelling.

1.2 State of research

1.2.1 Measuring soil erosion at the catchment scale

Soil erosion rates have been measured with different techniques for different purposes and on different scales. Starting with manual sampling and separately quantifying soil loss from field and sediment transport in streams, the need, interest and emphasis later shifted in science and society towards integrating catchments, considering physical and chemical properties of sediments, and studying single events with automated sampling (Olive and Rieger 1992; Walling et al. 2011).

Soil erosion at the catchment scale is typically derived from fluvial sediment transport and the sedimentation rate in reservoirs or lakes. Due to the high variability of soil erosion, long-term records are needed in order to assess trends, to capture rare events, and to obtain reliable average loads. Although sediment loads of rivers have been measured for a long time, questionable techniques, changed sampling strategies and sites often limit the use of early measurements (Walling 1997). Complementary long-term data is obtained from sedimentation rates in lakes and reservoirs (e.g., Zolitschka 1998; Brazier 2004; Rose et al. 2011). Both data types are used in this dissertation thesis to evaluate empirical erosion models.

Sedimentation rates from reservoir, pond, or lake data require reliable information on the trap efficiency and bulk density of sediment (Verstraeten and Poesen 2000, 2001; Verstraeten et al. 2003). After Habersack et al. (2008), the important questions for the monitoring of fluvial (suspended-) solids are: How representative is the sampling? How are the samples taken and analysed (filtered)? How often are samples taken? Direct and indirect (e.g., optical and acoustical) techniques have been applied at single and multiple points along the cross profiles of rivers to measure the fluvial transport of sediment (explained extensively in Bechteler 2006; Habersack et al. 2008; Kerschbaumsteiner 2010). The monitoring usually comprises the suspended fraction but rarely the bed load (Turowski et al. 2010). The frequency and timing of sampling is of great importance for the estimation of sediment loads and is, thus, an important source of uncertainty (Dickinson 1981; Walling and Webb 1981; Webb et al. 1997; Phillips et al. 1999; Ward 2008; Duvert et al. 2011) (see section 1.2.3). As direct sampling is laborious and limited, relationships between sediment load and water discharge (rating curves) (Walling 1977; Asselman 2000; Ward 2008) as well as turbidity (Davies-Colley and Smith 2001; Mano et al. 2009; Navratil et al. 2011) have been regularly used to interpolate and extrapolate sediment measurements. However, relationships between water discharge and sediment load and sediment concentration are highly variable and chaotic (Williams 1989; Sivakumar 2002; Sivakumar and Jayawardena 2003). They have to be carefully applied to avoid miscalculating sediment loads (Walling 1977; Walling and Webb 1981; Ward 2008).

Fallout radionuclides provide a complementary opportunity to obtain soil erosion rates in catchments at very different time scales (extensively reviewed by Mabit et al. 2008). Artificial Cesium-137 (^{137}Cs), cosmogenic Beryllium-7 (^{7}Be), and geogenic lead-210 (^{210}Pb) isotopes are strongly bound to soil particles. The radioactivity compared to undisturbed reference sites allows quantifying soil loss and deposition. King et al. (2005), Vrieling (2006) and in part Croft et al. (2012) reviewed remote sensing techniques and their application to map soil erosion features, to measure changes over time, and to derive controlling factors and model parameters.

The measured sediment yields of river catchments often aggregate several natural and anthropogenic, erosive and non-erosive sediment sources like phytoplankton and industrial effluents as well as transport processes in space and time. For erosion models, however, sediment data for calibration and validation should be appropriate in terms of scale, source and process. There has been a growing interest in quantifying the contribution and variability of sediment sources and linking erosion processes to sediment yield (Slaymaker 2003; Hoffmann et al. 2010; Walling et al. 2011). Monitoring, tracing with radionuclides, and source fingerprinting with sediment characteristics have been used to establish sediment budgets (Collins and Walling 2002; Walling 2006; Walling and Collins 2008; Walling et al. 2011; Hinderer 2012).

Although un-mixing models have also been proposed based on fingerprinting results (e.g., Collins et al. 1997; Fox and Papanicolaou 2008), typically only the total values measured at the catchment outlet are available. Under such circumstances, sources and processes have often been attributed to the variable relationships between (suspended-) sediment concentration and water discharge during storm events (Williams 1989; Smith and Dragovich 2009; Oeurng et al. 2010). Very few authors tried to directly separate the base load (i.e. the non-erosive sources) from total load and estimated the erosion-related or critical load as the difference between both. Walling and Webb (1982) and Bača (2008) linearly interpolated between pre- and post-event concentration. Hamm and Glassmann (1995) and Behrendt et al. (1999) derived annual and multi-annual average loads from average concentration and loads of discharge intervals.

1.2.2 Modelling soil erosion

Models always simplify the reality and have thus to be designed and selected for certain purposes (Wainwright and Mulligan 2004; Jakeman et al. 2006). Numerous erosion models have been developed over decades because of the complexity and scale-dependency of soil erosion and for a wide range of purposes (Merritt et al. 2003; Jetten and Favis-Mortlock 2006; Karydas et al. 2012). These models have, for instance, been used as management tools (Strauss et al. 2007; Kovacs et al. 2012) and for the quantification of diffuse nutrient fluxes in river catchments (Arnold et al. 1998; Prasuhn and Mohni 2003; Halbfaß et al. 2009; Venohr et al. 2011). Like other environmental models, erosion models can be roughly divided into the categories “empirical or statistical”, “conceptual”, and “physics-based” according to process detail, process understanding, and data demand (e.g., Merritt et al. 2003; Wainwright and Mulligan 2004). During the last few decades, there has been a shift from empirical relationships to more complex, process-oriented models in order to improve model predictions (Govers 2011). However, these complex models were shown not to be superior *per se* due to the huge spatio-temporal variability of soil erosion processes and controlling factors making parameterisation demanding as well as the inherent uncertainty and the problem of equifinality (Beven 1996; Jetten et al. 2003; Cherry et al. 2008; Govers 2011; Nearing and Hairsine 2011) (section 1.2.3 below). Accordingly, empirical models to predict sediment yield and soil loss in river catchments still prevail.

Many regression models have been proposed to estimate sediment yield from simple catchment parameters (as reviewed by Schäuble (2005) and de Vente et al. (2007)). Another common approach on the catchment scale is to estimate soil losses with the empirical universal soil loss equation (USLE, Eq. 1) and its derivatives (Auerswald 2008). While the first approach is not restricted to specific erosion processes and sources, the USLE and most other soil loss models only consider sheet and rill erosion (Merritt et al. 2003), i.e. in-stream erosion, sediment deposition, and other processes relevant in river catchments are neglected.

In the prominent USLE, the soil loss per unit area (E) is the product of factors which integrate environmental conditions and human activity: rainfall (R), soil erodibility (K), land cover and land use (C), erosion protection

(P), and topography (L·S) (Wischmeier and Smith 1978). Originally developed from extensive measurements on erosion plots in the U.S.A. to predict long-term average soil loss for conservation planning on agricultural land (Wischmeier and Smith 1978; Renard et al. 2011), the USLE has later been adopted to other regions like Germany (Schwertmann et al. 1987), purposes as cited in Renard et al. (2011), and scales (e.g., Kinnell 2010). USLE components are also used in other models (Nicks 1998).

$$E = R \cdot K \cdot C \cdot P \cdot L \cdot S \qquad \text{Eq. 1}$$

USLE estimates have to be corrected for deposition in order to compare them to sediment yields. Typically, models rely on the combination of USLE and sediment delivery ratios (SDR), a simple concept which has received criticism from Parsons et al. (2006) and others. However, the explicit modelling of transport processes suffers from the general limitations of complex models mentioned above. In its simplest form, a SDR is the black-box model of all the transport processes within catchments: the ratio between what leaves the catchment (i.e. the sediment yield) to what has been mobilised (soil loss E). Plenty of empirical relationships have been proposed in different regions to extrapolate SDR of river catchments (Walling 1983; de Vente et al. 2007). Alternatively, grey-box models aim to shed some light into the black box due to the internal variability of the sediment delivery (previous section). Local and travel-path characteristics, hydrology, and hydraulic connectivity have been used to disaggregate SDR (e.g., Ferro and Minacapilli 1995; Morehead et al. 2003; Lu et al. 2005). However, distributed models are rarely compared to the erosion pattern within catchments and may perform poorly at the raster-cell resolution despite being acceptable at the catchment scale (Jetten et al. 2003)

1.2.3 Assessing uncertainty and sensitivity in modelling

Model results are necessarily uncertain because models simplify and abstract reality. The term “model uncertainty” has many dimensions and sources (Tab. 3). According to Walker et al. (2003), uncertainty broadly arises from the “... lack of knowledge and ... [the] variability inherent to the system under consideration...” (p. 8). Uncertainty assessments are good scientific practice in modelling and should be communicated to model users (Walker et al. 2003; Caminiti 2004; Jakeman et al. 2006; Pappenberger and Beven 2006; Refsgaard et al. 2007). Many approaches and tools have been developed to assess model uncertainty and the appropriate methodology should be selected according to purpose, ambition, and model stage (Refsgaard et al. 2007). Sensitivity analyses are used to test the robustness of models and to assess the contribution of uncertainty sources on the model outcome in order to understand and reduce model uncertainty (Walker et al. 2003). In their recent review, Beven and Brazier (2011) state that only a few studies have aimed at uncertainty assessments and that models are often applied without them.

Given the multiplicative nature of the USLE and SDR models, the propagation of uncertainty is straightforward (Beven and Brazier 2011, p. 53). Some aspects of uncertainty are briefly reviewed below. The review is broadened in the following chapters.

Tab. 3: Dimensions and sources of model uncertainty after Walker et al. (2003)

Location	Level	Nature
Context (e.g., natural, economic, social)	Statistical uncertainty	Epistemic uncertainty
Model (structure, technical model)	Scenario uncertainty	Variability uncertainty
Inputs (driving forces, system data)	Recognised ignorance	
Parameters		
Model outcomes		

USLE estimates are very sensitive to topography, i.e. the L·S factor (Renard and Ferreira 1993; Risse et al. 1993), although the sensitivity depends on the value range (Auerswald 1987). Biesemans et al. (2000) used the Monte Carlo error propagation technique and found that the L·S factor contributed most to soil loss uncertainty. Likewise, Tetzlaff et al. (2013) identified the L·S and K factors as most contributing to the overall uncertainty of 34% in USLE estimations in the German federal state of Hessen. Their findings comply with values estimated for the federal state of Baden-Württemberg (Gündra et al. 1995).

Many studies have assessed sources of uncertainty in topographic parameters including resolution (scale), type of digital elevation model (DEM), and algorithm choice for topographic parameters (as reviewed by Wechsler 2007). Although the role of scale has been of great interest, only a few studies have addressed the spatial variability of topographic parameters is qualitatively affected on large scales. Kumar et al. (2000), Wolock and McCabe (2000), and Yong et al. (2009) compared a 100m- and 1000m-DEM and found a strong similarity for two common topographic parameters (slope and terrain index) calculated as mean values of 0.1°·0.1° to 1°·1° blocks in North America and China, despite significant quantitative differences. For 14 catchments of reservoirs in Spain, de Vente et al. (2009) similarly observed that a higher DEM resolution does not imply better predictions of soil erosion. While total sediment yields were equally well predicted with DEM of 30m and 90m resolution, the spatial pattern of soil erosion was even less reliable with the 30m-DEM.

When soil loss is estimated on regional or continental scales, another source of uncertainty arises from how to obtain maps of the USLE factors from limited input data. For most USLE factors, alternative approaches have been developed and applied. For instance, Panagos et al. (2012) recently extrapolated point data on measured K factors while van der Knijff et al. (2000) relied on the texture classes of the European Soil Database.

The uncertainty in sediment data is related to where, when, what, how, and how long has been measured (Tab. 4, section 1.2.1). The importance of sampling frequency and period is closely connected to the temporal variation of sediment transport in rivers. Olive and Rieger (1992) provide some early quantitative estimations of uncertainty in average annual sediment loads due to the sampling period. The standard error of the mean was typically >20% and long sampling periods are necessary to obtain more reliable estimates. Such long time series are also necessary to capture the contribution of few strong events. For instance, González-Hidalgo et al. (2009) found that 1% of the events contribute up to 30% of suspended-sediment loads in U.S. rivers.

Tab. 4: Sources of uncertainty in water quality data incl. suspended sediments after Rode and Suhr (2007)

Field instruments	Sampling location	Representative sampling	Laboratory analysis	Load calculation
Instrument errors	Point source inputs	Sampling volume	Sample conservation	Sampling frequency
Calibration errors	Impoundments etc.	Sampling duration	Sample transport	Sampling period
	Mixing of large tributaries	Spatio-temporal variation	Instrument error, Lab-induced uncertainty	Data extrapolation and interpolation

If data interpolation or extrapolation has to be applied due to low sampling frequency, the precision (range) and accuracy (bias) of load estimations can vary considerable. In an extensive study with European and U.S. suspended-sediment flux data, Moatar et al. (2006) found that the required sampling frequency for reliable results varies with flux regime and basin size. For a deviation below 20% from reference fluxes and a bias below 2%, sampling intervals range from less than 3 days for catchments smaller than 10,000 km^2 and with a high importance of single events (>40% annual suspended-sediment flux in 2% of time) to 20 days for basins larger than 200,000 km^2 where single events are less important. Similar results were reported by Coynel et al. (2004) and Horowitz (2003). Phillips et al. (1999) compared 22 extra- and interpolation methods from weekly to monthly sampling intervals in two river catchments. They found that the precision of any method significantly declines with sampling frequency with deviations often well above 100% for monthly sampling. Furthermore, their algorithm choice had a great impact on accuracy as the median deviation to the reference loads ranged from -79 to 5% and from -91 to -22% for weekly samplings. Webb et al. (1997) compared 7 methods and observed deviations between -49 and 222% for a simulated weekly sampling in another river catchment. In a short review, Vanmaercke et al. (2011) list studies which quantified other important sources of uncertainty: using log-transformed data for load estimation without bias correction (up to 50%), measuring at single instead of multiple points in the cross section (20%), the unknown fraction of bed load (30–50% for sand-bed rivers) and estimating the bulk density to convert sediment volume in reservoirs (0.17–1.7 Mg·m^{-3}). Navratil et al. (2011) assessed the global uncertainty in turbidity measurements to indirectly estimate suspended-sediment transport and the contribution of 9 uncertainty components. For a small mountainous catchment, they found an uncertainty of about ±30% for the annual load and of 20–50% for individual floods.

1.3 Aims and research questions

The dissertation thesis pursues two major research goals. First, the uncertainty and sensitivity in parameterising and evaluating empirical SDR models are assessed on the regional scale and beyond. For this purpose, SDR models in combination with the USLE are applied to predict annual to multi-year average sediment yields of river catchments. Second, the applicability of this modelling framework is evaluated in the European context and promising steps to improve the model performance and to overcome some restrictions are discussed. The motivation is to complement studies on validating European soil loss maps (van Rompaey et al. 2003b), analysing the relationships between soil loss and sediment yield in European catchments (Vanmaercke et al. 2012a), predicting the spatio-temporal variability of SDR and sediment yield in European regions (e.g., de Vente et al. 2007; Diodato and Grauso 2009), and assessing the uncertainty and sensitivity of soil erosion models and topographic indices (e.g., Gündra et al. 1995; Wolock and McCabe 2000; Yong et al. 2009; Tetzlaff et al. 2013).

The goal is to provide new insights into i) the relevance of topographic uncertainties and the uncertainty in critical yields of total suspended solids for the application, calibration and validation of empirical erosion models, ii) the sensitivity of SDR models to the estimation of USLE factors and iii) the pan-European applicability of empirical SDR models. The emphasis is put on systematically examining a broad range of parameters used for the USLE and SDR models, the variety of existing input data and algorithms to estimate model parameters, the catchment scale, and Europe-wide model applications. How the modellers' choices affect the model quality, i.e. the explained variability of SDR and sediment yields, is of special relevance because quantitative differences can be diminished during model calibration.

The alternatives for the assessment are selected from three categories which are of general relevance: topography, soil loss, and sediment data. Unfortunately, the multitude of available DEM and algorithms does not allow consideration of all the possible relationships between sediment data and model outcomes. To take into account as many factors as possible, the assessment has to be restricted to a few alternatives for each one.

First, topography and topographic parameters are fundamental for soil erosion and environmental models in general. Different DEM are available even for large-scale applications and many approaches have been proposed in the past to derive simple and complex parameters from these DEM. Plenty of studies have already been conducted on this topic. However, they mostly focused on a few topographic parameters as well as on the distribution and spatial pattern of raster values. A broad and systematic evaluation is still missing for semi-distributed erosion models, including the question which DEM is most suitable.

Second, each SDR and SDR model *per definitionem* depends on a soil loss map. As mentioned above, the USLE and its derivatives are the most common soil loss models for large-scale applications. However, the huge variability of soil loss and USLE factors has led to different approximations of USLE factors from limited input data. This raises the questions how to derive the USLE factors and soil loss maps and how these decisions influence model predictions.

Third, any SDR (model) also inherently depends on the sediment yield. In fact, any quantification of soil erosion at the catchment scale needs sediment data for calibration and validation. As mentioned above, estimating the sediment export from river catchments is associated with uncertainty and error. In the case of USLE-based models, long-term average annual sediment yields are needed which implies that the monitoring strategy is usually not subject to the modellers' decision. Even if we ignore the inherent uncertainty in a given set of actual measurements, the decision how to interpolate and extrapolate measured data strongly affects quantitative estimates of annual yields, as was pointed out in the previous section. Still, the consequences of algorithm choice for the variability in space and time have remained unclear until now. Another common problem is that the monitoring covers (filterable) suspended solids instead of sediments. Relating soil losses to observed loads of suspended solids overestimates the contribution of soil erosion (neglecting the bed load which is rarely systematically measured) if non-erosive sources like phytoplankton and industrial discharges contribute significantly. The inverse disaggregation of total suspended-solids data is not straightforward because of the huge variability of soil erosion and sediment fluxes. The few examples in the literature suggest two separate types of approaches to derive calibration and evaluation data for erosion models to be compared: statistical-lumped (e.g., Behrendt et al. 1999) and event-based approaches (e.g., Walling and Webb 1982). To analyse uncertainties and sensitivities in model evaluation and explore methodical limitations, the approach of Behrendt et al. (1999) was exemplarily compared to a new event-based approach based on the "functional streamflow disaggregation" (FSD) (Carl 2009).

Strictly speaking, neither the coarse input data nor the simplified modelling framework allow for proof that any erosion estimate is "better" than the alternative(s). The evaluation rather provides insights in how suitable the solution is for successfully establishing and applying SDR models and how strongly the choice affects the (explained) spatial and temporal variability of soil erosion at the catchment scale.

The European context is important for testing the applicability of erosion (SDR) models and working out regional limitations because the large range of environmental conditions and human activity correspond to highly variable sediment yields in Europe (Vanmaercke et al. 2011). However, such a large-scale and inter-regional assessment requires homogeneous input data for estimating the soil loss and SDR parameters. Unfortunately, the content and geometric resolutions of pan-European data are insufficient to directly calculate USLE factors. To accomplish the research goals, researching European sediment data and reviewing regional approximations of USLE factors have been important steps.

The following research questions and objectives are specifically addressed:

1. Are better resolved data helpful to improve the modelling of sediment delivery at the catchment scale? This is of special relevance for topography because of the huge range of available DEM which differ in resolution and content. In addition to the few studies on scale relationships, a wider range of DEM resolutions as well as simple and complex parameters related to soil loss and SDR estimations are included.
2. How do methods and input data influence calibrated models and the model evaluation? If the model evaluation is sensitive to these choices which alternative has to be recommended?
   Which approaches have been proposed in the past to estimate USLE factors from European data? Are alternatives relevant for the evaluation and applicability of SDR models? Do the numerous approaches to estimate topographic parameters from DEM or alternative estimations of erosion-related fractions of total suspended solids (for model calibration and validation) play a role at the catchment scale, i.e. are careful choices promising to improve the predicted soil erosion or has the model to be adjusted? Specifically, does the higher working resolution of the FSD approach allow disaggregating more reasonable annual critical yields than the statistical approaches?
3. Which parameters improve the prediction of the spatial and temporal pattern of sediment yields and sediment delivery ratios? Which application constraints exist for the empirical SDR model?
   Due to the high variability of soil erosion, sediment yields of catchments with different environmental characteristics and long sampling periods are required to adequately evaluate model results. Are there general catchment and data properties which explain high residuals and depict model limitations?
4. How uncertain are model results because of alternative data and algorithms within the given empirical modelling framework and the given data base? Which choices contribute most to the model uncertainty?

This dissertation thesis is part of the evaluation and development of MONERIS (http://www.moneris.igb-berlin.de), a conceptual, semi-distributed model (Behrendt et al. 1999; Venohr et al. 2011). MONERIS is applied to quantify and apportion nutrient fluxes in large river catchments primarily inside but also outside of Europe. MONERIS and similar models recognize soil erosion as an important diffuse pathway of nutrients into surface waters and implement empirical relationship to estimate the variability of SDR. In MONERIS, the modelled SDR is combined with the long-term average soil loss estimated with the USLE and European data to obtain average sediment yields. These values are weighted with annual rainfall to get annual yields. MONERIS was calibrated with critical yields of total suspended solids. Although MONERIS is seldom directly addressed in the following chapters, the results are expected to be relevant for better understanding how predictions of sediment-bound nutrient fluxes in European river catchments can be improved with this (and similar) models.

1.4 Outline of the dissertation thesis

The following chapters 2–5 of this cumulative dissertation thesis have either been published in or submitted to peer-reviewed journals. For better coherence and legibility, cross references have been adjusted and citation styles, abbreviations, units and symbols harmonised. The readability of some figures and tables has also been improved. Chapter 6 summarizes the main findings followed by a brief outlook in chapter 7.

The research has been conducted in two study areas. For the regional analyses in chapters 2–4, catchments of monitoring gauges situated in two German federal states have been chosen to cover a broad range of area and terrain. For some of these gauges, multi-annual time series of daily data on water discharge and suspended-solids concentrations have been available for model evaluation. This regional study is the basis for the European context envisaged in chapter 5.

Chapter 2 addresses the research questions 1–2. From various DEM, simple and complex topographic parameters are derived with alternative algorithms. Non-topographic parameters of erosion models are not considered. The alternative solutions therefore depict the relative influence of topographic uncertainty on applications of erosion models (and many other environmental models). The uncertainty in model parameterisation and modelled soil erosion is quantified followed by correlation analyses to evaluate impacts on the spatial pattern. The contribution to this study is 100% my own.

In chapter 3, the analyses of the previous chapter are broadened to the estimation of sediment yields. Similar to chapter 2, alternative model solutions are compared after varying the estimation of topographic parameters. Additionally, algorithm choices to estimate annual and long-term average critical yields are considered. Long-term average annual sediment yields of river catchments are modelled by combining the USLE (Eq. 1) and a spatially disaggregated SDR. Simple regression models are tested to explain the annual variability. In preparation of chapter 5, model parameters are derived from pan-European data. Model results are evaluated using other calibrated empirical models and the estimated critical yields. Catchment properties are assessed as potential model parameters to improve model predictions. Model limitations are also discussed. This study addresses the research questions 1–4 in which the contribution to it is 95% my own.

Chapter 4 deals with the inter-annual variability of sediment yields and is an extension of chapter 3. A new conceptual model is applied to one catchment for which a multi-decadal set of annual suspended-solids yields has been available. In this study, the regression models used in chapter 3 are compared to the new model based on seasonally weighted rainfall data. The improved model predictions, the model applicability in Europe, and model limitations are discussed. The study addresses the research question 3. I contributed to about 33% of the study concept, the literature review, data analyses, discussion and writing and to about 75% of the data research and processing.

In chapter 5, the parameters of the improved SDR model in chapter 3 are used for an empirical SDR model to explain the spatial variability of SDR and sediment yields of European catchments. Prior to the model application, sediment yields measured in rivers and reservoirs had to be collected. In this study, alternative empirical approaches to approximate USLE factors from pan-European data are applied to quantify the uncertainty in the modelled soil loss and sediment yields. The uncertainty is compared to the uncertainty in sediment data. In a sensitivity analysis, the impacts on the spatial pattern of soil loss and SDR, model predictions and the applicability of such an empirical modelling framework are evaluated. Model and data constraints are discussed. The applicability and sensitivity of the SDR model is compared to other empirical SDR models. This study addresses the research questions 2–4 and the contribution is 100% my own.

2 Topographic uncertainty and catchment-based models

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2.1 Abstract

Topographic attributes are key parameters in numerous models to assess sediment or nutrient input into surface waters. A broad range of DEM and algorithms bring, however, uncertainty to topographic interpretations. This may raise the question whether empirical, semi-distributed models can cope with such uncertainty. In this study, primary and complex topographic attributes related to soil loss and distributed sediment delivery were computed from DEM with cell widths between 10 and 1,000 m. Correlation and regression analyses were conducted with average values of the catchments of 138 German gauges spanning different terrain. Two slope, single-flow routing and slope-length algorithms were also included to evaluate their effects. Although either choice mostly induces significant changes of catchment means of slope, flow length, slope-length factor and SDR, Spearman's rank correlation coefficients are generally above 0.9. It is suggested from the data that linear or slightly curved functions are suitable to adapt average topographic attributes computed from differently resolved DEM or by different methods. Empirical catchment-based models can thus cope with topographic uncertainty and model users may implement these equations to compare model outputs. However, the catchment delineation and stream definition may constrain their application.

2.2 Introduction

Topography is a fundamental controlling factor for many processes within the landscape (Moore et al. 1991). Consequently, primary topographic attributes such as slope and specific catchment area (SCA) or more complex indices have been used in numerous environmental models. However, the wide range of DEM resolutions, their inherent accuracy and the algorithms applied to compute topographic attributes are sources of uncertainty in any topographic analysis (Wechsler 2007).

The underlying processes of soil erosion and sediment transport are complex and spatially heterogeneous thus limiting the application of physics-based models to comparatively small areas (de Vente and Poesen 2005; Lenhart et al. 2005). Empirical approaches are therefore common in large-scale models. The USLE (Eq. 1) is widely applied to assess soil loss within catchments and its slope-steepness (S) and slope-length (L) factors reflect the influence of terrain. However, most mobilised soil particles are deposited along the transportation path and do not reach the outlet. In empirical models, the relationship between observed sediment yield (SY) at the outlet and modelled gross erosion is called sediment delivery ratio (SDR) and various relationships between the SDR and simple catchment parameters such as area or average slope have been proposed (Walling 1983). The complex relationships and the cumbersome prediction of SY using the catchment area alone are extensively discussed in de Vente et al. (2007). Distributed approaches implement therefore environmental information like land use or topography to spatially disaggregate the SDR (Veith 2002).

Many studies have been conducted to evaluate scale, cell size or algorithmic impacts on topographic attributes. Either choice will significantly alter the derived values and may influence the outcome of subsequent models (Wechsler 2007). The impact thereby depends on terrain complexity. Previous studies have mostly focused on cell-based statistics (Wu et al. 2005; Wilson et al. 2007; Wu et al. 2008) and spatial pattern (Tarboton 1997; Endreny and Wood 2003). However, semi-distributed or lumped models use average values for (sub-) catchments. Only few studies have specifically assessed cell size effects on catchment means of topographic attributes. Comparing average values between a 100m- and a derived 1000m-DEM in 1°·1° blocks spatially scattered across the U.S.A., linear relationships for slope β, SCA and the topographic index defined as ln(SCA/tan β) are proposed in Wolock and McCabe (2000). Testing areas across China, a more recent study supports these results by covering more pronounced terrain and by comparing a 100m- and an independent 1000m-DEM (Yong et al. 2009).

Based on these findings, the main problem addressed in this study is: Can empirical catchment-based sediment or nutrient input models cope with uncertainty in terrain representation? This leads to the questions:

a) Can catchment means of topographic attributes be converted between different DEM resolutions?

b) Is this, in addition to Wolock and McCabe (2000) and Yong et al. (2009), also possible for complex parameters related to soil erosion and sediment delivery and for DEM resolutions below 100 metres?

c) Are results of common GIS algorithms correlated as well?

2.3 Methods

2.3.1 Study area and input data

The study area consists of the catchments of 138 gauges in North-Rhine Westphalia (NRW, West Germany) and Bavaria (South Germany). They span various terrains from lowland to alpine conditions as well as a wide range of areas between 20 and 8,800 km^2 (median of 326 km^2) and are partly nested. Official gauge positions and corresponding catchment areas have been provided by the Bavarian State Office for the Environment and the NRW State Agency for Nature, Environment and Consumer Production.

Tab. 5 lists the available DEM with their horizontal and vertical resolutions. Since DEM correction compromises topographic parameters (Callow et al. 2007), these DEM were pre-processed as little as possible. Even projections were left unchanged, except for DEM100 whose geographic projection was transformed to UTM using bilinear interpolation. However, bridges are abundant in DEM10 and DEM50 and had to be eliminated prior to any assessment. All GIS operations were performed with ESRI ArcGIS 9.2 (Esri 2006).

Tab. 5: Specifications of DEM

Name	Resolution XY (m) / H (m)	Coverage	Source
DEM10	10 / 0.01	Eastern NRW	NRW Surveying and Mapping Agency
DEM50	50 / 0.1	NRW	NRW Surveying and Mapping Agency
DEM100	100 / 1	Germany[a]	SRTM (Jarvis et al. 2006)
DEM250	250 / 1	Germany	Federal Agency for Cartography and Geodesy
DEM1000	1,000 / 1	Germany[a]	GTOPO30 (available from the U.S. Geological Survey)

[a] And adjacent areas

2.3.2 Bridge removal

Bridges are artificial barriers impeding correctly modelled water flow paths. High-resolution ATKIS polygon data on land use and land cover in 2007 (Germany Survey, NRW) was used to remove obstacles in water bodies after a visual comparison revealed its geometrical agreement with DEM10 and DEM50. ATKIS classes like rivers, wetlands and weirs were reclassified as streams and a local rectangular minimum filter was applied to the elevation of stream cells. The expected maximum width of bridges determined the neighbourhood size. It had to be large enough to contain at least one water cell for each barrier cell.

In mountainous areas, however, roads span valleys that are not characterised as water bodies in ATKIS. So (and if no suitable dataset was available), a simple topographic approach was developed to approximate continuous flow paths. At first, raster cells with a catchment area above 5 km^2 in a minimum-filtered DEM were considered as stream cells. This arbitrary threshold was chosen to limit the removal of potential flow barriers to well-established streams. These preliminary streams were then iteratively expanded to cover neighbour cells below or at the same height in the original DEM. Expanding and shrinking by half of the minimum filter size connected the separate segments and led to the stream mask. All raster cells lower or equal to the local average height were considered as belonging to such a segment. Each cell within the mask higher than a threshold above the local minimum was eventually classified as barrier cells to be filtered. A centreline approximation (THIN function) was implemented to exclude shoreline cells.

2.3.3 DEM processing

After filling sinks, flow direction and upslope area were calculated for each DEM. As a change of DEM resolution or flow routing algorithm also changes upslope areas of raster cells, gauges then had to be manually adjusted to cells of high flow accumulation. Raster cells comprising an upslope area of at least 2 km^2 were defined as stream cells after a visual comparison with a river net of Germany (BKG 2003). These stream cells were thus considered as to be congruent with the river network. If possible, the spatial relations of gauges were also taken into consideration.

Grid-based flow routing algorithms behave differently in distributing the outflow of raster cells to downslope neighbours. The simplest and widely-applied approach is to follow the steepest descent along one of the eight cardinal directions in a raster (D8 algorithm). Disadvantages are the poor spatial congruence (Endreny and Wood 2003) and the SCA distribution with a high proportion of raster cells having low values (Wilson et al. 2007). Therefore, the D∞ (D-infinity) approach (Tarboton 1997) was also included. It expresses the flow direction as a continuous angle between 0 and 2π and thus enables the flow to be diverted to a maximum of two neighbour cells. D∞ functionality was provided by the freely available TauDEM extension for ArcGIS (Tarboton 2005). An own flow accumulation routine was developed to circumvent problems with floating-point numbers and for automation purposes.

Tab. 6: Parameters derived from DEM

Topographic parameter	Methodology
Catchment area A	Flow accumulation including outlet cell
Flow length to outlet and stream	ArcGIS flowlength function (D8 flow only, TauDEM D8 flow direction recoded to ArcGIS scheme)
Sediment delivery ratio	Veith (2002) but without land use factor and dj=0.9997 / 30 m for stream cells
Sediment transport capacity index (STI)	$STI = \left(\frac{SCA}{22.13}\right)^{0.6} \cdot \left(\frac{\sin\,\beta}{0.0896}\right)^{1.3}$ (Moore and Wilson 1992)
Slope β	Neighbourhood (β_{Nbh}) (ArcGIS) and maximum slope method (β_{max}) (ArcGIS, TauDEM)
Slope length factor L	van Remortel et al. (2001), with cutoff slope angles of 0.7 (β<5%) and 0.5 (β≥5%) (D8 flow only, TauDEM D8 flow direction recoded to ArcGIS scheme)
Slope steepness factor S	Nearing (1997)

ArcGIS and TauDEM furthermore implement different slope algorithms. As this may be relevant for soil erosion and sediment input modelling (Warren et al. 2004), the sensitivity to the choice of methodology was compared against DEM resolution effects. Due to technical constraints, TauDEM could not be applied to DEM10.

2.3.4 Statistics

Statistics about each topographic parameter listed in Tab. 6 were calculated for all catchments whose areas deviated less than 25% from DEM100 (slope) or official values (other statistics). Spearman's rank correlation coefficients and regression equations were then determined to assess the relationships between the datasets. Additionally, Wilcoxon tests were applied to estimate the significance of differences. Although STI and LS are only meaningful for land areas, the stream delineation may vary according to the chosen approach (upper catchment area threshold, SCA-slope relationship among others). Therefore, statistics were computed with and without stream cells to obtain a general idea of dependencies on stream cell definition. All statistical analyses were performed using the R-based software Statistical Lab v3.7 (FUB 2008).

2.4 Results and discussion

2.4.1 Catchment area and stream delineation

After bridge removal and manual adjustment of the gauges, the median of the ratios between modelled and official catchment areas is close to 1.0 for all DEM and all flow algorithms. Inter-quartile ranges and the number of outliers, however, increase with raster cell size (Fig. 2). Nonetheless, individual errors occur for every DEM and do not follow any clear trend.

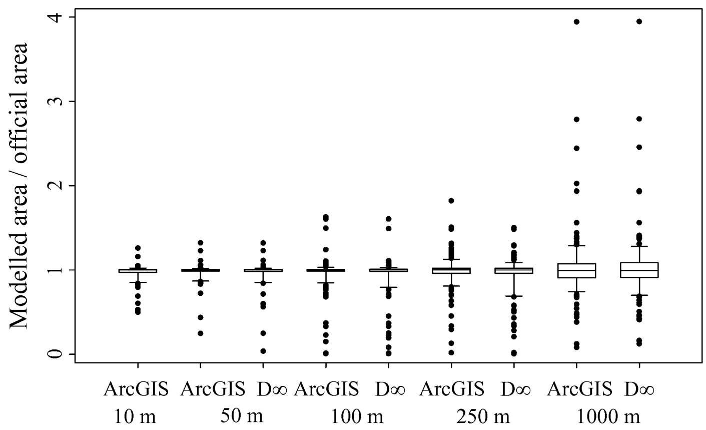

Fig. 2: Modelled area to official area for different DEM resolutions and flow routing algorithms

The catchment area of each raster cell obviously depends on DEM resolution as larger raster cells also accumulate more area. Consequently, with a given area threshold, the proportion of stream cells in catchments is expected to be proportional to cell width (see Fig. 4). Nevertheless, flow routing algorithms can also have a considerable impact (Fig. 5). The D∞ algorithm of TauDEM returns more stream cells than ArcGIS for DEM1000 (median ratio 1.37). This is not only a bifurcation effect but partly a result of flow routing in flat terrain as

TauDEM's D8 algorithm returns streams which are also 19% longer. Such areas are more prominent in the smooth surface of DEM1000 (section 2.4.2). The differences for the other DEM are small and found to be significant for DEM50 ($p<0.001$) and DEM100 ($p=0.02$ for D∞, $p=0.06$ for D8). They are not significant for DEM250.

By contrast to other topographic attributes, correlation coefficients are not only relatively low between DEM resolutions (Tab. 7) but also between flow routing algorithms. For the latter, r_S values depend on DEM resolution being lowest for DEM1000 with $r_S \approx 0.6$ (ArcGIS), $r_S<0.5$ (TauDEM) and increasing to $r_S \approx 0.8$ for all other DEM. The statistical relationships are linear.

2.4.2 Slope and flow length

The highly significant increase ($p<0.001$) of average slope angles with DEM resolution can also be seen in Tab. 7. With raster cell sizes becoming smaller, average slopes also converge (Fig. 3 left). This pattern is similar for both slope algorithms, although the impact is slightly lower for the maximum (D∞) than for the neighbourhood slope algorithm (Fig. 4). Correlation coefficients exceeding 0.9 in all cases support the inter-scale correlation of average slopes observed by Wolock and McCabe (2000) and Yong et al. (2009). Linear regression models describe well the relationships, especially between fine DEM resolutions. Seven virtual gauges with catchment means of slope angles between 10% and 35% have been included in the statistics to increase the sample size for DEM10 and DEM50.

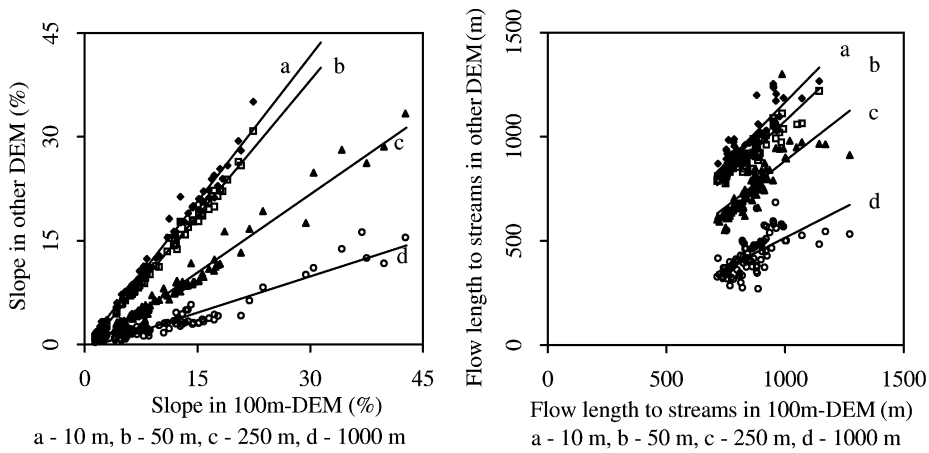

Fig. 3: Average slope and average flow length to streams in relation to DEM100, left: slope (β_{Nbh}), right: flow length (ArcGIS)

Differences between slope algorithms are comparatively small and proportional to cell width (Fig. 5). They are significant ($p<0.001$) for all DEM besides DEM10. The D∞ algorithm returns the highest slope angles. This is a flow routing effect because maximum slopes computed with both D8 algorithms do not differ significantly apart from DEM250 (p=0.03). The respective coefficients of linear regression models are 1.00 (0.99 for DEM250).

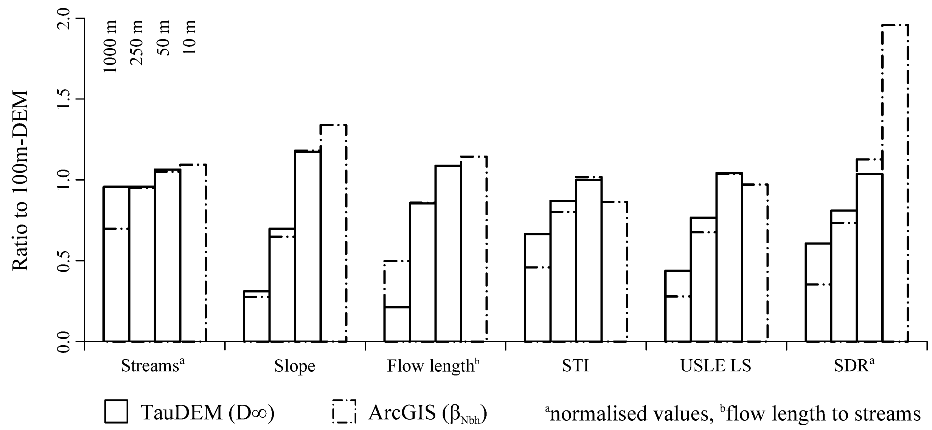

Fig. 4: Effect of DEM resolution on stream delineation and average topographic attributes

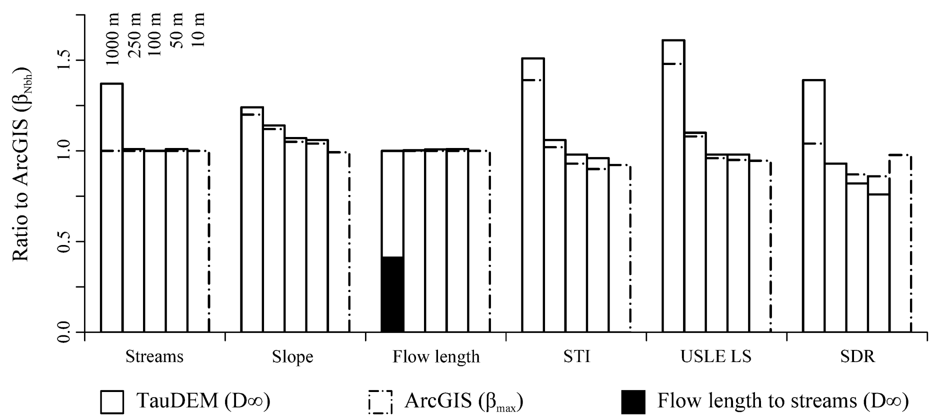

Fig. 5: Method effects on stream delineation and average values of topographic attributes

Average flow lengths prove to be reciprocal to DEM cell size (Tab. 7) because flow paths meander more if grid cells become smaller. However, the deviations are not as high as for slope angles. The relationships for flow lengths to the outlet are linear and correlation coefficients exceed 0.98. Flow lengths to streams in DEM10 to DEM250 behave in a similar manner, although the impact is larger and r_S values are slightly smaller. In contrast to these DEM, flow lengths in DEM1000 are not only considerably shorter (Fig. 3 right) but they are also moderately correlated to other DEM resolutions. These correlation coefficients are reciprocal to the proportion of stream cells (Tab. 7; r_S=0.65 for TauDEM D8).

Tab. 7: Coefficients of regression equations $y = a_2x^2 + a_1x$ respectively $y = a_1x$ with x = DEM100

		Neighbourhood slope				Maximum slope (D∞)		
Parameter	DEM	10	50	250	1,000	50	250	1,000
Streams	a_1	0.11	0.52	2.38	6.89	0.53	2.39	9.60
	n	59	63	112	91	61	108	91
	r_S	0.79	0.90	0.79	0.66	0.80	0.83	0.49
Slope	a_1	1.37	1.23	0.70	0.31	1.20	0.73	0.37
	n	66	71	113	94	61	108	91
	r_S	0.98	0.99	0.98	0.93	0.99	0.98	0.94
Flow length to	a_1	1.11 (1.16)	1.03 (1.08)	0.87 (0.88)	0.83 (0.51)	1.04 (1.09)	0.87 (0.88)	0.83 (0.21)
outlet	n	59	63	112	91	61	108	91
(to streams[a])	r_S	0.98 (0.85)	0.98 (0.92)	0.99 (0.93)	0.99 (0.73)	0.98 (0.94)	1.00 (0.95)	1.00 (0.51)
STI[a]	a_2	0.00	0.00	0.01	0.01	0.00	0.00	0.01
(STI)	a_1	0.88 (0.76)	1.05 (1.11)	0.78 (0.82)	0.35 (0.27)	1.02 (1.01)	0.94 (0.92)	0.59 (0.51)
	n	59	63	112	91	61	108	91
	r_S	0.99 (0.94)	0.99 (0.99)	0.99 (0.99)	0.93 (0.93)	1.00 (1.00)	0.99 (0.99)	0.96 (0.97)
USLE LS[a]	a_2	0.00	0.00	0.01	0.01	0.00	0.00	0.01
(USLE LS)	a_1	1.02 (1.05)	1.09 (1.11)	0.70 (0.67)	0.21 (0.19)	1.08 (1.10)	0.90 (0.86)	0.38 (0.27)
	n	59	63	112	91	61	108	91
	r_S	0.99 (0.99)	0.99 (0.99)	0.99 (0.99)	0.93 (0.94)	0.99 (0.99)	0.99 (0.99)	0.95 (0.96)
SDR[a] (SDR[b])	a_1	0.20 (0.20)	0.56 (0.23)	1.82 (2.80)	3.43 (13.92)	0.50 (0.11)	2.12 (2.59)	5.68 (22.23)
	n	59	63	112	91	61	108	91
	r_S	0.71 (0.85)	0.97 (0.62)	0.97 (0.80)	0.87 (0.72)	0.95 (0.63)	0.92 (0.69)	0.82 (0.69)

[a] Non-stream cells, [b] TauDEM D8

The choice of methodology plays a marginal role for average slope and flow length to outlet. Flow length values are only slightly higher in TauDEM- than in ArcGIS-processed DEM. Differences are reciprocal to DEM resolution and statistically significant ($p<0.001$) except for DEM1000. Correlation coefficients are above 0.99 for both parameters and linear regression models fit well (Tab. 8). In contrast, the impact is considerable for DEM1000 when flow lengths are calculated to the streams (Fig. 5). In accordance to section 2.4.1, the correlations are only moderate (Tab. 8; r_S=0.76 between ArcGIS and TauDEM D8). Although the slight differences for all other DEM are also significant ($p<0.001$, $p=0.005$ for DEM50 and D∞), r_S values are higher.

Tab. 8: Coefficients of regression equations y = a_1x with x = ArcGIS (β_{Nbh})

		ArcGIS (β_{max})					TauDEM (D∞)			
Parameter	DEM	10	50	100	250	1,000	50	100	250	1,000
Slope	a_1	0.99	1.03	1.02	1.08	1.21	1.05	1.04	1.10	1.23
Flow length	a_1	1.00	1.00	1.00	1.00	1.00	1.02	1.01	1.00	1.00
(to streams)		(1.00)	(1.00)	(1.00)	(1.00)	(1.00)	(0.99)	(0.98)	(0.99)	(0.40[c])
STI[a] (STI)	a_1	0.93	0.90	0.93	0.97	1.38	0.95	1.00	1.00	1.45
		(0.75)	(0.62)	(0.76)	(0.71)	(0.73)	(0.65)	(0.81)	(0.73)	(0.75)
USLE LS[a] (USLE LS)	a_1	0.95	0.95	0.92	0.91	1.46	0.97	0.94	0.91	1.50
		(0.95)	(0.94)	(0.91)	(0.88)	(1.12)	(0.96)	(0.93)	(0.88)	(1.14)
SDR[a] (SDR[b])	a_1	0.98	0.86	0.87	0.93	1.04	0.76	0.82	0.93	1.39
		(0.83)	(0.53[d])	(0.52)	(0.43)	(0.68)	(0.24)	(0.52[d])	(0.45[d])	(0.69)

[a] Non-stream cells, [b] TauDEM (D8), [c] r_S=0.55, [d] 0.85<r_S<0.90

2.4.3 Sediment transport capacity index (STI) and USLE LS factor

Besides the slope angle, three factors influence the dependency of both attributes on cell size changes. At first, the DEM resolution determines the minimum possible erosive slope length and SCA. Secondly, the terrain-smoothing effect of coarse DEM resolution (section 2.4.2) can further increase slope length values. Finally, average slope angles decrease and so does the exponent of the L factor.

The inter-scale correlation of SCA values (Wolock and McCabe 2000; Yong et al. 2009) is supported by the high correlation of average STI values between DEM (Tab. 7). However, the relationships between coarse and fine DEM are not linear but slightly curved. In accordance to the equivalence of STI and the LS factor (Moore and Wilson 1992), both attributes show a similar pattern of resolution dependency, although the impact on LS is higher than on STI (Fig. 6 left and centre). They are also highly correlated (r_S>0.99) and second-order polynomial regression models describe the relationship between average LS and STI values.

The largest average values occur in DEM50, nonetheless DEM100 differs only slightly (p>0.03). The reason for the observed drop of average STI and LS in DEM10 has to be the decrease of both SCA and erosive slope length prevailing higher average slopes. This corresponds with observations by Sørensen and Seibert (2007) on a cell-by-cell basis. In addition, Fig. 4 shows that values calculated by the neighbourhood slope method are stronger influenced by DEM resolution changes, although differences to the maximum slope method decline with raster cell size. The r_S values between DEM are high (Tab. 7). However, calculated values and differences are somewhat extreme because neither land use patterns nor reasonable slope length caps were applied.

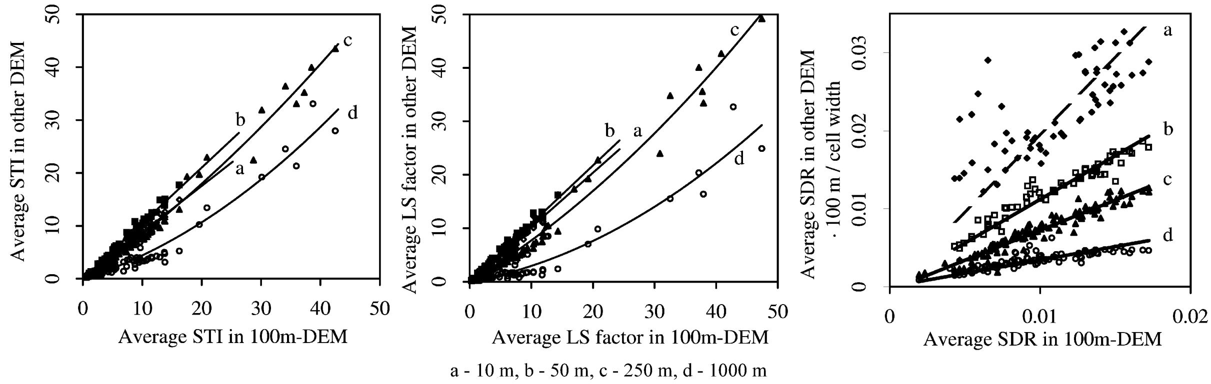

Fig. 6: Average STI, LS, and normalised average SDR for non-stream cells in relation to DEM100, all topographic calculated with neighbourhood-slope method (β_{Nbh})

Changing methods has a significant ($p<0.001$) influence on average STI and LS values aside from DEM1000 (STI) and DEM250 (LS) between both D8 algorithms. Following the pattern for slope angles, the effect is proportional to DEM resolution (Fig. 5). While the maximum slope methods (D8 and D∞) return higher STI and LS values for DEM1000, the effect declines and even reverses for finer resolved DEM. Correlation coefficients are also very high ($r_S>0.98$) and linear regression models describe the observed relationships (Tab. 8).

Stream cell definition affects both topographic attributes differently. The equations in Tab. 8 and Tab. 7 for the LS factor generally change a little when stream cells are included in the statistics. Only in DEM1000 with its many stream cells, the inter-algorithmic relationships are affected (Tab. 8). Besides, the impact on STI is much higher because stream cells have *per definitionem* large SCA values resulting in high STI values if slope angles $\beta>0°$. This is especially relevant for the neighbourhood slope method which may induce high slope angles along narrow valleys or non-corrected bridge cells. Consequently, the difference is comparatively large and the correlation low for DEM10 (Tab. 7). The relationship between neighbourhood and maximum slope changes considerably for all DEM, particularly for DEM1000 (Tab. 8). Nonetheless, correlation coefficients remain high ($r_S>0.98$; $r_S=0.95$ for DEM10).

2.4.4 Sediment delivery ratio (SDR)

Modifying the DEM resolution or GIS algorithms does not affect many lumped SDR approaches provided that the catchment area does not change. If appropriate, the empirical relationships for average slope may be applied to transform SDR values to other DEM resolutions or methods. Given the high correlations for the STI and USLE LS factor, the sediment input could also be estimated.

By contrast, the distributed algorithm is inherently resolution-dependent as the number of raster cells along each flow path increases with cell sizes becoming smaller. Therefore, the impact of changing DEM resolution is highest among topographic attributes (Fig. 4, $p<0.001$). If means are corrected for cell width, they are proportional to average slope. Correlation coefficients between DEM are low in comparison to other topographic parameters (Tab. 7). The moderate values for DEM10 ($0.59<r_S<0.71$ for ArcGIS β_{Nbh}; $0.52<r_S<0.64$ for ArcGIS β_{max}) indicate an information content that cannot be fully explained by coarser DEM. The relationships are linear (Fig. 6 right).

Average SDR values also change significantly if flow routing or slope calculation is altered (Fig. 5). Mostly, p is below 0.001 except for DEM10 ($p=0.01$) and the D8 comparison for DEM250 ($p=0.003$). At a first glance, the results may contradict the slope pattern. However, higher average maximum slope angles do not mean higher values for each raster cell. At the feet of slopes, for example, the neighbourhood slope algorithm considers the upslope area whereas the maximum slope only includes the flatter downslope relief. Due to the shorter distance to either streams or outlet, the slope angle at a foot of slope has a higher significance than the slope itself where maximum slope returns higher values. Terrain characteristics as well as DEM resolution determine thus the spatial pattern and absolute values of slope differences and, eventually, empirical relationships. The considerably larger average SDR for the TauDEM-processed DEM1000 is equivalent to the larger number of stream cells (section 2.4.1).

Without stream delineation, the raster cells next to the outlet are most important and maximum slopes are mostly gentle here. The neighbourhood method can consider (steeper) slopes. This is more probable in coarse DEM, given the fact that the differences between regression equations with and without stream cells are higher there (Tab. 8). However, the opposed trend of surface smoothing seems to partly overlay. Correlation coefficients are slightly lower between both DEM resolutions and algorithms. The relationships still follow a linear trend although the coefficients change noticeably (Tab. 7 and Tab. 8).

2.5 Conclusions

Previous studies suggested that average slope angles and SCA computed from a 100m-DEM can be downscaled to 1,000 m resolution using linear equations (Wolock and McCabe 2000; Yong et al. 2009). Supporting these findings, the results show that linear or slightly curved regression models are principally suitable to transform catchment means of topographic attributes for DEM resolutions between 10 metres and 1,000 metres. This is not only possible for simple but also for complex parameters such as USLE's slope-length factor or spatially distributed SDR. Furthermore, linear equations describe sufficiently the relations between outcomes of different single-flow routing and slope algorithms. This is important as often only one version is implemented in standard GIS software. Despite significant changes in absolute values, catchment-based models are thus capable to handle topographic uncertainty. However, there are three constraints.

Firstly, the relations are only valid as long as catchment delineation is not compromised. If catchment areas have to be derived from a DEM, higher resolution means better conformance with official values. Discrepancies here will also affect many lumped SDR approaches. Nevertheless, if DEM become too detailed artefacts may emerge interfering with the calculation of flow directions and catchment area. Secondly, the regression models for the spatially distributed SDR and the STI seem to be susceptible to stream definition. However, the two tested alternatives of a constant threshold of upper catchment area for all DEM resolutions and no streams, respectively, give only a general impression. Finally, the inter-resolution correlations are comparatively low for the distributed SDR approach. The moderate correlations between DEM10 and the other DEM indicate a content being not fully explicable.

DEM and method choice significantly influences estimates of sediment or nutrient input to surface waters. Although it is feasible to expect better results with higher DEM resolution, this study shows that this may not be true for empirical, semi-distributed models. Pre-processing and computing time does not only increase rapidly but benefits of better resolved DEM or elaborate algorithms are marginal. Model users may use the proposed empirical equations to compare model results based on different DEM resolutions or topographic algorithms. Nonetheless, an overall sensitivity analysis of sediment and nutrient models has to include other relevant factors such as land use or soil.

3 Improving the estimation of erosion-related suspended solid yields in mountainous, non-alpine river catchments

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3.1 Abstract

Sediment yields in river catchments may be modelled using a range of DEM and data algorithms. We assessed the impact of alternative choices on the evaluation of empirical modelling approaches. The modelling framework in this study consisted of the universal soil loss equation, a published local SDR, and commonly available input data. The study area comprised the catchments of 31 monitoring gauges for which daily data of suspended solids (SS) was available. For these stations, we studied the effects of two interpolation schemes for daily SS concentrations and of two approaches to separate erosion-related ("critical") fractions from total SS yields (SY) on model evaluation. Despite a good agreement between modelled and critical SY, the unexplained spatial variability was considerable. For additional 109 catchments, we quantified impacts of two DEM resolutions, two slope algorithms, and three slope length algorithms on modelled SY. DEM resolution and slope algorithm proved to be most relevant for the model uncertainty. High correlation coefficients between the respective alternatives revealed the minor relevance of data or algorithm choices for model quality. Correlation analyses showed that the SDR model lacked a hydrological parameter. Adjusting the modelled SDR accordingly significantly increased the explained variability of SY.

3.2 Introduction

Human activity has significantly accelerated the natural processes of soil erosion and sediment relocation causing global concern (Quinton et al. 2010). For surface waters, the delivery of sediments poses serious threats to aquatic habitats and water quality, particularly because of the nutrients and pollutants attached to the sediments (Bilotta and Brazier 2008).

In order to mitigate diffuse riverine pollution and achieve sustainable river management, estimates of sediment inputs are therefore critical. However, the underlying processes are highly variable in space and time (González-Hidalgo et al. 2009), making monitoring in situ both complex and costly. Consequently, numerous modelling approaches have been proposed (as reviewed by Merritt et al. 2003). With increasing computing power and capabilities of geographical information systems (GIS), spatially distributed models have become more able to predict runoff, soil loss and (suspended) sediment yields in river catchments. However, the performance of complex, physically-based models is not necessarily better than those of empirical models, due to process variability, parameterisation and equifinality (Jetten et al. 2003). Furthermore, the inherent variability in replicated erosion measurements (e.g., Boix-Fayos et al. 2006) and sediment observations made Foster et al. (2001) emphasize that it is more important to find comparative, relative estimates, than quantitatively accurate ones. For these reasons, the empirical USLE is still widely applied (Auerswald 2008), despite its conceptual limitations (Novotny and Chesters 1989) (Eq. 1).

Generally, only a small fraction of mobilized soil particles reaches the outlet, and therefore, USLE estimates have to be adjusted with deposition. In the simplest form, catchment-wide SDR are applied to the USLE as correction factors. The spatial variability of such SDR has often been explained in terms of simple catchment properties such as area or average slope (de Vente et al. 2007). This "black-box" concept has received criticism (e.g., Novotny and Chesters 1989); however, the alternative explicit modelling of sediment flow is cumbersome, and limited to small areas (de Vente et al. 2007). As a compromise, for larger catchments, "grey-box" models spatially disaggregate the SDR by considering connectivity (Dymond et al. 2010) or such features as channel distance, flow path properties, land use, and topography (e.g., Fu et al. 2010).

In this study, a modelling framework consisting of the USLE, a disaggregated SDR and (mostly) pan-European data was used to estimate annual suspended-sediment yields for topographically diverse river catchments. The main goal of our study was to identify the most relevant factor to improve the predictions of the empirical modelling framework. We followed the argument of Foster et al. (2001) and aimed at a relative agreement between model results and validation data.

In this context, we quantified the implications of several data and method choices and their relevance for (improving) the quality of model predictions. Assessing such choices is of interest because a variety of available input data and a broad range of algorithms exist to derive model parameters, especially for topography (e.g., Warren et al. 2004; Vaze et al. 2010). Moreover, validation data also depends on method choices (e.g., Phillips et al. 1999). We have focussed on three choices: i) interpolation algorithms of monitoring data of suspended solids (SS), ii) algorithms to estimate the erosion-related fraction from monitoring data, iii) topographic algorithms and the DEM resolution to derive topographic parameters for the model.

The remainder of this chapter proceeds as follows. Section 3.3 briefly describes the study area. Section 3.4 presents the methods for the interpolation of SS data, the estimation of erosion-related fractions, the parameterisation of the USLE and the SDR including the effect of topographic representation on both model parts, and the identification of catchment properties controlling the spatial and inter-annual variability of SS yields. Section 3.5 discusses the quantitative and qualitative consequences of data and algorithm choices for modelling sediment yields and model evaluation. After presenting the controlling factors of SS variability, there is a proposal for an improved SDR model. Chapter 3 ends with conclusions for catchment-based sediment studies.

3.3 Study area

The study area comprised catchments of 140 gauges in the German states of Bavaria and Nordrhein-Westfalen (NRW) (Fig. 7). These partly nested catchments also extend into other German and Austrian federal states and small parts of the Czech Republic. The catchment areas calculated from DEM were between 19 and 25,600 km^2, with a median of 385 km^2. These values agreed well with those officially reported by the State authorities.

The Bavarian gauges are located in the basins of the Danube and river Main, a tributary of the Rhine. The Danube divides the Alps and the Alpine Foothills in the south from the eastern highlands of the Bayerischer Wald and the Fränkische Alb. The NRW catchments are situated in the North German Plain, in the basins of the rivers Rhine, Weser and Ems. This area changes from flat, glacially sculpted landscapes in the north of NRW to the mountainous region in the south and east.

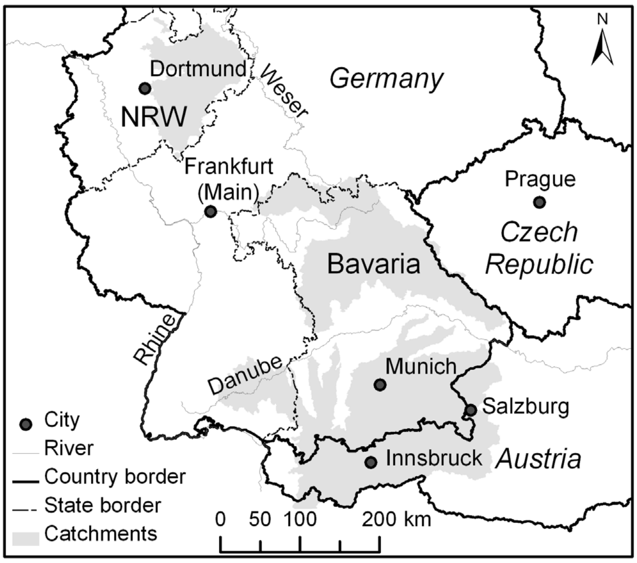

Fig. 7: Location of the study area (NRW – Nordrhein-Westfalen)

The climate alters from maritime in the west to more continental in the east. The calculated mean annual precipitation in the study area of 1,070 mm·a^{-1} is above the typical German range from 500 mm·a^{-1} to 800 mm·a^{-1} (BMU 2003). The precipitation gradient from west to east is superimposed by topography. Annual values exceed 1,000 mm in highlands and 2,000 mm in the Alps.

Human activity has shaped most of the catchments and their water bodies. There are flood protection measures and water diversion (river Altmühl), river straightening along the river Ilz, and hydropower plants along the rivers Salzach and Isar. In NRW, mine subsidence areas are of special relevance to hydrology.

Land use and land-use change in Bavaria and NRW are typical for Germany. Land-use distribution largely depends on topography. Arable land is usually abundant in lowlands, and changes to pasture in highlands, and to open spaces in the Alps. Since 1970, agricultural land and forests have covered about 50% and 30% of the total study area respectively. Most of the agricultural land has been arable land with a rising proportion over the time. Cereals are the dominant crop. In recent decades, maize has been increasingly cultivated, but root crops less so (Destatis 1971, 1980, 1990, 2009). The expansion of maize cultivation has been associated with higher sediment yields in the Austrian Danube basin (Summer et al. 1996). This observation is presumably also valid for the German areas.

Tab. 9: Distribution of management strategies on arable land in 2009/10 to mitigate soil losses (Destatis 2011a; b; Chamber of Agriculture of Upper Austria, pers. comm.)

Federal state	Winter soil cover %	Intermediate crops %	Conservation tillage %	No-till %
Baden-Württemberg	70	20	40	1
Bavaria	60	25	20	1
NRW	80	15	30	1
Upper Austria	70[a]	n.a.	15[b]	n.a.

[a] Estimated, 35% winter grain + 35% soil cover financed by the agro-environmental programme ÖPUL, [b] Including no-till farming

The good agricultural practice in Germany requires a soil management that reduces soil losses, especially from arable land. The most common management strategy in the study area is to cover soils during winter (Tab. 9). Less relevant measures can, however, be of regional importance. For instance, intermediate crops are grown on almost half of the arable land in the mountainous south of Bavaria (Bavarian State Office for Statistics and Data Processing, pers. comm.) and mulching is common for maize and in the Tertiärhügelland (Tertiary Hills) where the risk of soil loss is highest (Bavarian State Research Centre for Agriculture, pers. comm.). In NRW, winter soil cover and conservation tillage are most abundant in the south-east (IT.NRW State Office for Information and Technology, pers. comm.). Comparable data was not available for Austria. Nonetheless, arable land is scarce in the Austrian sub-catchments and soil losses from there are less important for sediment yields.

3.4 Methods

3.4.1 Introduction

The modelling framework consisted of two parts. First, a combination of USLE and distributed SDR was applied to estimate long-term average sediment yields. We evaluated how well the model explained the spatial variability within the study area. Second, hydrological data was related to suspended-solid yields to explain the inter-annual SS variability.

To describe the implications of data and method choices for the model outcome and model quality, we correlated the alternative long-term averages among all catchments to assess the spatial variability and alternative annual values for each catchment to assess the temporal variability. We assumed that a high correlation coefficient indicates that the choice is of less importance for the explained variability, i.e. a poor model performance cannot be attributed to an unfavourable choice but to limitations of the USLE or the SDR model. For this case, we used correlation analyses to identify catchment properties to improve model predictions. In contrast, a low correlation coefficient implies that the model performance can be improved by a careful choice.

The estimation of erosion-related fractions, the modelling framework, and identification of controlling factors and the statistical analyses are presented in this chapter. Statistical analyses were performed using the software package R 2.10.1 (R Development Core Team 2009).

3.4.2 Estimation of critical yields of suspended solids (SS)

3.4.2.1 Monitoring data on suspended solids

Daily values of SS concentration (SSC) and water discharge (Q) for 31 monitoring stations were provided by the Bavarian Environment Agency (Table 2). SSC and Q have mostly been monitored for 20–30 years. The time series of SSC were interpolated from single point measurements taken with buckets at irregular intervals. The sampling frequency depended on flow conditions and typically ranged from multiple samples per day to one sample a week. Multi point measurements were occasionally carried out to calibrate and validate the standard routine (all information from LfU 2000).

The Bavarian Environment Agency has adopted two different methods (henceforth method A and B) to interpolate daily means from the sampling data. In method A, the concentration SSC_i measured or linearly interpolated (Eq. 2) at time t_i (in hours) was time-weighted using the trapezoidal rule (Eq. 3).

$$SSC_i = \frac{SSC_{i-1} + SSC_{i+1}}{Q_{i-1} + Q_{i+1}} \cdot Q_i \qquad \text{Eq. 2}$$

$$SSC_A = \frac{1}{24} \sum_{i=1}^{n} \left(0.5 \cdot \left[SSC_i + SSC_{i+1}\right] \cdot \left[t_{i+1} - t_i\right]\right) \qquad \text{Eq. 3}$$

n is the number of trapezoids covering one day (SSC values at midnight were also linearly interpolated). In method B, the trapezoidal rule was similarly applied to interpolate SS loads as product of SSC_i and Q_i instead of

SSC_i. If one sampling was taken at a day, Q_i was the daily mean value. Otherwise, Q_i was interpolated at time t_i (in seconds). SSC_B was the ratio of daily average SS load and daily Q.

Occasionally there were small gaps in the SSC data, which we filled by linear interpolation (Q ≈ const.) or estimated from the sampling data taken at these days. For most monitoring stations, the results of both interpolation alternatives were available, thus we assessed the consequences of method choice on suspended solid yields for model evaluation.

Mean specific runoff q of the gauges and intra-annual dynamics were closely related to topography (Tab. 10). Accordingly, the rivers were categorized into non-alpine (lowlands and highlands) and alpine. Alpine rivers were typically found to have peaks of runoff and sediment yield during late spring and summer because of the combined effect of melting snow and summer precipitation. In contrast, most non-alpine rivers had their minimum of q and SS yields (SY) in these seasons due to increased evapotranspiration. However, a few catchments did not clearly fit into either group. The intra-annual dynamics were extenuated by lakes, dams and barrages for the rivers Mindel, Amper and Isar.

3.4.2.2 Estimation of critical SS yields

Monitoring gauges necessarily integrate all SS sources and processes within their catchments. Although soil erosion is the main source of suspended solids in Bavaria (Habersack et al. 2008), comparing model outcomes to total yields (e.g., Auerswald 1992) would generally overestimate its contribution. In order to evaluate an erosion-based model, it is necessary to estimate the erosion-related fraction from an observed total SY. This genuine inverse task is challenging because no simple relationship exists between SS dynamics and other processes within rivers and river catchments (Sivakumar and Jayawardena 2003).

Assuming that non-eroded fractions (e.g., phytoplankton, industrial effluents) are almost independent of water discharge, the erosion-related (henceforth "critical") fraction was estimated for each gauge using long-term average SS concentrations and loads for discharge classes following Behrendt et al. (1999) (Fig. 8). In most cases, SSC and loads increased considerably above a gauge-specific critical discharge Q_{crit}. To obtain Q_{crit}, Q was first quantized using varying bin widths. For each bin, the average values for Q, SSC and SS load were calculated and linear regression models were developed for low Q (RM_{low}) and high Q (RM_{hi}) to predict SSC and SS loads from Q. Q_{crit} was determined as the intersection of RM_{low} and RM_{hi}. Critical yields are the difference between RM_{hi} and RM_{low} for $Q>Q_{crit}$. In other words, RM_{low} is the base load SY_{base} not related to soil erosion. In this study, SY_{graph} refers to the critical yields of this graphical-statistical approach.

Tab. 10: Overview of gauges for validation, collection periods and basic catchment properties, H – average height, β – average slope, Ar – arable land (Corine Land Cover 2000), Pr – interpolated long-term average annual precipitation (section 3.4.3.3.1), and q – measured average annual runoff

Gauge	River	Area	H	β[d]	Ar	Pr	q	Availability
		km²	m	%	%	$mm \cdot a^{-1}$	$mm \cdot a^{-1}$	(q and SY)
Aha	Altmühl	693	453	4.0	55	781	192	1976–1999
Bad Kissingen	Fränk. Saale	1,572	382	8.1	53	820	267	1974–2003
Burghausen[e]	Salzach	6,649	1,268	36.8	3	1,422	1,205	1970–2002
Coburg	Itz	346	455	10.6	40	979	466	1991–2002
Dietldorf	Vils	1,100	465	7.3	39	843	308	1988–2003
Duggendorf	Naab	5,434	504	7.3	38	854	299	1971–2003
Gemünden	Sinn	610	431	14.2	13	991	401	1984–2003
Harburg	Wörnitz	1,569	475	5.2	58	856	247	1988–2003
Hohenkammer	Glonn	390	504	4.0	73	965	263	1971–2003
Hüttendorf	Regnitz	3,864	422	6.2	40	805	253	1989–2003
Inkofen	Amper	3,076	622	6.1	37	1,132	489	1971–2003
Kalteneck	Ilz	760	714	13.0	8	1,209	682	1971–2003
Kemmern	Main	4,224	433	9.3	44	891	336	1971–2003
Kempten[e]	Iller	955	1,182	30.4	4	1,754	1,539	1971–2003
Laufermühle	Aisch	956	342	5.4	57	738	171	1971–2003
Mainleus	Main	1,166	481	9.1	43	932	408	1993–2003
Muggendorf	Wiesent	660	465	7.8	54	934	344	1971–2003
Neumühle	Rednitz	1,847	424	5.3	42	775	226	1990–2001
Oberaudorf[e]	Inn	9,715	1,809	45.7	1	1,240[c]	985	1971–2003
Offingen	Mindel	949	597	4.8	27	1,040	426	1970–2003
Pettstadt	Regnitz	6,991	404	6.4	46	812	246	1971–2003
Plattling[a]	Isar	8,435	741	13.5	34	1,199	658	1970–2003
Regenstauf[b]	Regen	2,658	593	12.2	22	1051	463	1972–1991
Ruhstorf	Rott	1,052	447	5.5	76	933	287	1971–2003
Schärding[e]	Inn	25,664	1,204	30.2	15	1,218	903	1971–2003
Schlehenmühle[b]	Roter Main	71	488	6.1	33	904	269	1978–1992
Schönach[b]	Große Laber	407	419	4.9	72	790	196	1989–1992
Stein[be]	Traun	367	852	23.7	5	1,779	1,074	1971–1992
Taferlruck[ab]	Große Ohe	19	1,003	18.6	0	1,265	977	1983–1992
Unterjettenberg[be]	Saalach	927	1,213	42.8	0	1,693	1,262	1971–1992
Vilsbiburg[b]	Große Vils	320	484	4.9	81	946	282	1972–1992

[a] Intra-annual runoff pattern similar to alpine catchments, [b] Only SSC_B data available (section 3.4.2.1)
[c] Adjusted (section 3.4.3.3.1), [d] Neighbourhood approach, [e] Alpine catchment, [f] Calculated with Corine Land Cover (2000)

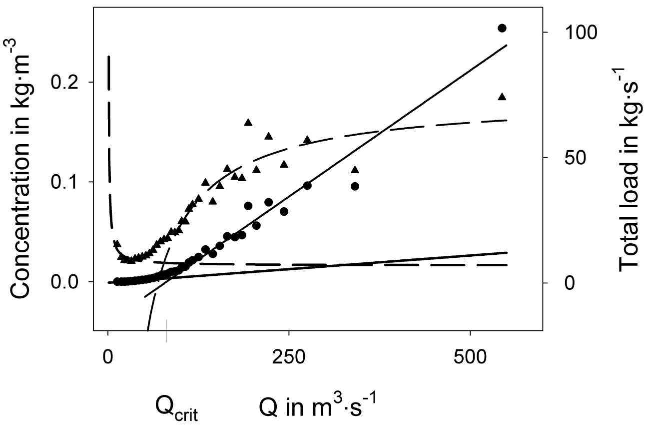

Fig. 8: Estimation of critical yields of suspended solids according to Behrendt et al. (1999). Average SS concentrations (triangles) and SS loads (circles) for Q intervals were used to establish regression models for high and low Q. The critical yield SY_{graph} for each gauge was derived from the difference between both regression models for SS concentrations (broken lines) and SS loads (straight lines).

SSC_B values (section 3.4.2.1) were used to compare SY_{graph} estimates to estimates obtained with the "functional streamflow disaggregation" (FSD) technique (Carl and Behrendt 2008). In FSD, total discharge is seen as superposition of conceptual runoff components. This data-sparing method was originally used to decompose time series of daily discharges from headwater catchments (Carl et al. 2008). The components were found to reasonably approximate classical overland flow, interflow, and base flow as generated by distributed hydrological models, though a range of uncertainty remained in these tentative assignments.

FSD code version 1.0 (Carl 2009) was used to estimate the overland flow Q_{fast} by the "fast" FSD component. Very small negative values were – as round-off artefacts (P. Carl, pers. comm.) – set to zero. For technical reasons, the time series were split into subsets of 11 years maximum. Possible boundary effects were avoided using an overlap of 2 years. SY_{FSD}, the alternative to SY_{graph}, was finally obtained as the product of q_{fast} and SSC_B with $q_{fast} = Q_{fast} \cdot A^{-1}$. The term "critical SY" subsumes SY_{graph} and SY_{FSD} throughout this study.

3.4.2.3 Evaluation of verification data

For both algorithms for interpolation and critical-yield estimation, we compared the spatial variability of long-term average critical SY and the annual values for single gauges. Histograms and Kolmogorov-Smirnov tests showed that the data was not normally distributed. Thus the non-parametric Spearman rank correlation coefficient (r_S) was calculated to assess qualitative differences. Additionally, Wilcoxon tests were conducted to test whether quantitative differences between alternatives were significant. The null hypothesis was rejected if $p \geq 0.05$. To obtain quantitative differences, we calculated the slope of the regression model $y = a_1x$.

To our knowledge, no established inverse method exists to derive critical yields from total SS yields. Validation was not possible but we modified the algorithm for SY_{graph} to assess the plausibility of results and to explore differences between critical SY. Three modifications were applied for gauges with i) high dissimilarity between average SY_{FSD} and SY_{graph}, ii) low correlation of annual values, and iii) large discrepancy to model results (section 3.3.5). First, bin widths for Q quantization were calculated from quantiles to assess the effect of alternative binning on SY_{graph} estimates (SY_{alt_bin}). Second, the alternative SY_{tot_base} was calculated as the difference between total SY and the regression model for low Q (SY_{base}) to overcome possible limitations of the linear regression model for high Q. Third, SY_{graph} was calculated for single years to better reflect the inter-annual variability of sediment yields (SY_{ann_graph}).

3.4.3 Modelling approach

3.4.3.1 Model outline

Sediment yields were modelled using raster datasets. The geometric resolution of the raster DEM was set as reference for all other input data. All GIS operations were performed with ArcGIS 9.2 (Esri 2006). The main steps consisted of i) defining streams as a prerequisite for the distributed sediment delivery ratio (SDR), ii) estimating the USLE factors and soil loss, and iii) estimating the SDR. For each non-stream raster cell, gross soil erosion and the SDR were calculated to estimate the amount of eroded soil that eventually reaches the stream network and the outlet. For stream cells, it was assumed that no soil loss occurs. Long-term in-stream deposition was also neglected because no information was available that would allow parameterisation. Similar to the USLE estimates, sediment yields (SY) as a product of SDR and USLE soil loss were long-term annual means.

3.4.3.2 Deriving the stream network

For defining the starting points of the stream network, we adopted the approach of Colombo et al. (2007). The study area was subdivided into 3 regions according to precipitation, lithology, relief, vegetation cover, and soils. For each region, we analysed the relationship between local slope (β) and specific catchment area to define the minimum catchment area for stream initiation. Raster cells with catchment areas between 1.1 and 1.4 km^2 were defined as starting points. Additionally, surface curvature was considered to include visible linear surface features like small rivers and valleys. It was assumed that surface runoff is dominantly linear here and thus the USLE is not valid.

The D8 algorithm as proposed by Jenson and Domingue (1988) was applied for flow routing and subsequent stream delineation from raster DEM. This algorithm was readily implemented in the ArcGIS software package.

It exclusively assigns the outflow of a raster cell to the raster cell along the steepest descent in its 3x3 neighbourhood. The streams were combined with water bodies taken from the Corine Land Cover 2000 dataset (CLC2000) (EEA 2007a). Artificial waterways were not considered.

3.4.3.3 Estimating soil erosion

USLE factors were estimated using coarse datasets and empirical approaches (Tab. 11). Coarse resolution does not only refer to geometry but to content as well. For instance, the European Soil Database (ESBN and EC 2004) contained 6 texture classes instead of specific percentages for silt, clay and sand. Therefore, all factors were estimated with empirical relationships which have been published for the study area or which are widely accepted. As several approaches have been suggested in the literature for the L factor, the consequences of choices were exemplarily assessed.

Tab. 11: Overview of methods and data to estimate USLE factors (β – slope angle)

Factor	Dataset	Data (Resolution)	Method	Reference (Method)
R	Long-term average precipitation (Pr, $mm \cdot a^{-1}$)	Points (Variable; 0.5° raster)	$R_{Bavaria} = 0.078 \cdot Pr + 4.3$ $R_{NRW} = 0.1096 \cdot Pr - 14.25$	Strauss and Blum (1994) DIN (2005)
K	European Soil Database	Polygons (1:1 Mio)	$K = 0.0086 \cdot silt\% + 0.0083$	Strauss et al. (2003)
C·P	CLC2000; agricultural statistics	Polygons (1:100,000; administrative units)	C = f(land use) and P = 0.85 for arable land C constants and P = 1.0	Kagerer and Auerswald (1997); Auerswald (2002); Stumpf and Auerswald (2006)
L·S	SRTM-DEM (Jarvis et al. 2006); DEM (NRW Surveying and Mapping Agency)	Raster (100 metres; 50 metres)	$S = -1.5 + \frac{17}{1 + e^{2.3 - 6.1 \cdot \sin\beta}}$ Iterative L factor L = f(β)	Nearing (1997) van Remortel et al. (2001) Asselman et al. (2003); Fuchs et al. (2010)

3.4.3.3.1 R factor

A raster dataset of long-term average annual precipitation (Pr) was interpolated from calibrated annual data obtained from the German Meteorological Service (DWD) and monthly data from the Global Precipitation Climatology Centre (GPCC), using an inverse distance weighting algorithm. To improve interpolation across the Alps and Austria, we included stations for which only long-term means were available (1961–1990, 1971–2000) (DWD, Austrian Central Institute for Meteorology and Geodynamics). The uncalibrated DWD data (Pr_{uncal})

was adjusted to altitude H (in metres) using a linear relationship (Eq. 4, Behrendt unpubl.). R factors were estimated from this dataset applying specific empirical relationships for NRW and Bavaria / Austria (Tab. 11).

$$Pr = 0.0525 \cdot H + 63 + Pr_{uncal} \qquad \text{Eq. 4}$$

The results of our approach were tested for plausibility because station density was low for Austria. Pr and runoff q fitted well for all catchments except the alpine Oberaudorf which lies largely in Austria. This was confirmed by an independent raster dataset (HISTALP) (Efthymiadis et al. 2006; ZAMG 2009). Therefore, we changed our Pr value of 1,028 $mm \cdot a^{-1}$ to the more realistic HISTALP value (Tab. 10).

3.4.3.3.2 K factor

For each soil typological unit, the soil database provided information on the dominant surface textural class (attribute TEXT1). The class value was translated to silt content according to van der Knijff et al. (2000).

3.4.3.3.3 C·P factor

According to Auerswald (2002), statistical information on land use was evaluated to determine the percentage of small grain and sod-forming crops for administrative units (Tab. 12). Information on the percentage of root crops in mulch-tillage systems was scarcely available. This value was set to 2.5% in NRW (Schmidt et al. 2007) and to 0% elsewhere. The C value was assigned to all arable land and heterogeneous agricultural areas within an administrative unit (CLC2000 codes 21x and 24x). These C factors were spatially disaggregated because C factors are i) much higher for arable land than for other land uses and thus more important for soil loss in river catchments and ii) largely dependent on crops and crop rotation schemes.

For some raster cells in Bavaria, we found a combination of arable land on very steep slope angles which exhibits a high risk of soil loss. We assumed this combination to be unlikely and to be the result of inaccuracies in the input data (e.g., the minimum mapping unit of CLC2000 of 25 ha is above the DEM resolution). Land use classes 21x and 24x were thus replaced by pasture (CLC class 231) on slope angles above 25% (Strauss 2007).

Tab. 12: Statistical data on land-use used to determine C factors for arable land

Region	Spatial domain	Year	Data source
Austria	Federal state	1999	Statistik Austria
Bavaria, Baden-Württemberg, Thüringen	County	2003	State Offices for Statistics
Czech Republic	Region (*kraj*)	2003	Czech Statistical Office
Nordrhein-Westfalen (NRW)	County	2004	Chamber of Agriculture

3.4.3.3.4 L·S factor

The SRTM-DEM was transformed to 100 metres raster resolution. Bridges which affected water flow routing and catchment delineation in flat terrain were removed from the 50m-DEM using a simple GIS-based approach (section 2.3.2).

Slope angles for the S factor and the SDR were computed with the maximum (β_{max}) and a neighbourhood method (β_{Nbh}). The latter is the standard plane-fit algorithm as implemented in the ArcGIS software package.

We compared three L factors. First, we used a regression model between β_{Nbh} (in °) and L factors measured in the German Federal State of Baden-Württemberg (Eq. 5) (Fuchs et al. 2010). Second, the L factor was derived from a constant erosive slope length of 100 metres following Asselman et al. (2003) and Liu et al. (2000) (Eq. 6). Third, we applied the iterative GIS algorithm proposed by van Remortel et al. (2001) (L_{GIS}). For L_{GIS}, erosive slope lengths were separately computed from slope changes along flow routes for CLC2000 classes. The upper limit was set to 3 raster cells (300 metres), the maximum of L_{emp}. For forests, a maximum of L_{GIS} = 2 was defined (Stumpf and Auerswald 2006).

$$L_{emp} = 0.0028 \cdot \beta_{Nbh}^{3} - 0.0937 \cdot \beta_{Nbh}^{2} + 0.729 \cdot \beta_{Nbh} + 1.3038 \qquad \text{Eq. 5}$$

$$L_{100m} = \left(100\,\text{m} \big/ 22.13\,\text{m} \right)^{\frac{b}{b+1}} \text{ with } b = \frac{\sin\beta}{0.0896 \cdot \left(3 \cdot \sin^{0.8}\beta + 0.56\right)} \qquad \text{Eq. 6}$$

3.4.3.4 *Estimating sediment yields (SY)*

Sediment delivery to streams was calculated for each raster cell as a product of long-term soil erosion and the sediment delivery ratio (SDR). A modified SDR approach of Halbfaß and Grunewald (2008) was applied, approximating the land use factor α with the C factor (Eqs. 7–8).

$$SY_i = \max\left\{ \overline{a}_L \cdot \sqrt{\overline{\beta}_L / L_i}\,, 1 \right\} \cdot E_i \qquad \text{Eq. 7}$$

$$\alpha_i = 1.43 \cdot \ln(C_i) + 9.49 \qquad \text{Eq. 8}$$

For each raster cell i, the original local land use factor α_i and slope β_i were replaced by the downslope averages $\overline{\alpha}_L$ and $\overline{\beta}_L$ to represent flow path characteristics. These averages were calculated by summing up all values α_i and β_i along the flow path and dividing the sum by the length L_i of the flow path.

All variables in the equation above refer to the centres of non-stream cells. Local SDR of raster cells next to stream cells below 5% indicated a low probability of sediment transfer into stream cells. The upslope areas of such cells and of urban areas were excluded.

3.4.3.5 Verifying soil erosion, sediment yields and sediment delivery ratios

Sediment sources other than sheet and rill erosion are *sensu stricto* outside the scope of the chosen model approach. This is certainly the case for alpine catchments where landslides and stream erosion are important processes (Tab. 10). Model evaluation was therefore only possible for non-alpine catchments. Among those catchments, total sediment yields measured at Schlehenmühle (Roter Main) and Kalteneck (Ilz) were extraordinary high (SY_{tot}=42 $Mg{\cdot}km^{-2}{\cdot}a^{-1}$). This is due to urban runoff at Schlehenmühle (Strohmeier et al. 2005) and an alpine-like character of the river Ilz and its tributaries (R. Brandhuber, pers. comm.). For all other catchments, former studies showed that the USLE is sufficient to estimate sediment yields.

Modelled soil loss was indirectly verified by applying the empirical SDR equations proposed by Auerswald (1992) and Behrendt et al. (1999) to the USLE results. Both SDR equations were originally calibrated with Bavarian SS monitoring data similar to our dataset. To test the agreement of modelled soil erosions, regression analyses were conducted. In accordance to the methods and data used in both studies, the products of SDR and USLE results were compared to total SS load and SY_{graph}.

Linear regression models ($y = a_1x$) were applied to evaluate the different modelled sediment yields and SDR. The slope a_1 was used as a measure for average quantitative differences to critical SY and SDR. The significance of a_1=1.0 (match) was obtained with t-tests. The adjusted r^2 were calculated as a measure of how well the models explain the quasi-observed variability of critical SY.

“Critical SDR” values were calculated as ratio of critical SY to modelled soil erosion. They were correlated to modelled SDR in order to assess whether the USLE or the SDR approach failed to capture the spatial variability in the study area. Similar to the verification of evaluation data (section 3.4.2.3), Spearman’s rank correlation was conducted for correlation analyses. Weak correlations between modelled and critical SY suggest that the empirical approaches for the USLE factors or the USLE itself were inappropriate while weak correlations for SDR indicate the need for improving the SDR approach.

3.4.3.6 Analysing topographic uncertainty

The regression models used for model evaluation were also used to quantify the topographic uncertainty. Correlation coefficients were evaluated to assess whether method choices or DEM resolution affect the explained variability of the verification data.

3.4.4 Improving model results

3.4.4.1 Variability of sediment yields (SY)

To complement the model structure, correlation analyses were conducted to identify catchment-wide controlling factors of critical-SY and SDR variability. Our reasons for this approach were two-fold. First, the available data resolution limits the cell-based SDR approach to land use and terrain. However, there are many more environmental and anthropogenic factors which influence hydraulic connectivity and sediment delivery. Second, the USLE only returns average annual values of soil loss. Nonetheless, it is well-known that SY and SSC can vary considerably over time, as a combination of variability in soil loss and sediment transport. A small number of events can contribute large fractions of annual yields, and their role may even be enhanced by reservoir flushing (e.g., Hamm and Glassmann (1995) for the river Salzach). However, their distribution and frequency may not be well related to annual precipitation used in the present study to estimate USLE R factors (Auerswald 1992).

Only a few studies in the Bavarian and Austrian river catchments have related environmental and anthropogenic factors such as geology, flow distance, water discharge (Bauer 1967), land use and land use change (Summer et al. 1996), power plants (Weiss 1996) to sediment variability. Auerswald (1992) analysed total SS for 22 Bavarian catchments and proposed a linear relationship to annual R factors, catchment area and soil loss. Behrendt et al. (1999) also developed a linear model for the inter-annual variability of SY.

3.4.4.2 Identifying factors controlling the variability of suspended solids

We tested the strength and significance of correlations between possible controlling factors (Tab. 13) and the set of critical SY and SDR. All but the climatic and hydrological factors were considered to be static. We analysed the spatial variability of long-term average SSC and SY among the catchments. For climatic and hydrological factors, annual values were calculated to analyse the temporal variability for each catchment. Because the SDR was defined as the ratio of SY and long-term mean soil erosion, the inter-annual variability of critical SY and critical SDR were identical. The statistical methods were similar to those mentioned in sections 3.4.2.3 and 3.4.3.5.

Tab. 13: Catchment properties (possible controlling factors) which were correlated to critical SY and SDR (see Restrepo et al. (2006) for definitions and references)

Factor group	Factors
Land use	Proportion of urban areas, arable land, pasture, forests and open spaces (%)
	Average distance of agricultural land to streams (l_{agri}, km)
Morphometry	Catchment area (A, km^2)
	Average slope (β, %)
	Average, minimum and maximum height (H, H_{min}, H_{max}, m)
	Catchment length (L_C, km)
	River length (L_R, km)
	Drainage density (DD, km·km^{-2})
	$(H-H_{min})/(H_{max}-H_{min})$ (HI, -)
	H/A (β_1, m·km^{-1})
	H/H_{max} (H_{pk}, -)
	H_{max}/L_C (Hr, m·km^{-1})
Erosion	E (Mg·km^{-2}·a^{-1}) and USLE factors
Climate and Hydrology	Annual and long-term average Q_{tot}, Q_{fast} (m^3·s^{-1})
	Annual and long-term average Q_{tot}/A (q_{tot}, mm·a^{-1}), q_{fast}
	Annual precipitation (Pr_{ann}, mm·a^{-1}) and Pr (mm·a^{-1})
	Pr_{ann}/Pr_M 1970–2003 (PP, -)
	Modified Fournier index 1970–2003 (MFI, mm·a^{-1})
	Annual and long-term average SSC (kg·m^{-3})
Mixed	β·C, H·Ar, H·Pr

Annual precipitation data was not available for all stations. Additional steps were thus necessary to estimate annual catchment means of precipitation (Pr_{ann}). They were calculated by first interpolating GPCC data and DWD stations for which annual sums of precipitation were available (Pr'_{ann}). Pr'_{ann} and its long-term mean Pr' was applied to derive Pr_{ann} from the long-term precipitation Pr already used for the USLE R factor (section 3.4.3.3.1) (Eq. 9). Similarly to Pr, we used the HISTALP dataset to test for plausibility. The inter-annual variability of both precipitation datasets was found to be consistent for each gauge ($r_S \approx 0.9$).

For the calculation of the Modified Fournier index (MFI) (Arnoldus 1980), average maximum monthly precipitation (Pr_M) was estimated similarly to Pr_{ann} by first interpolating monthly GPCC data ($Pr_{M,\,GPCC}$). $Pr_{M,\,GPCC}$ and the annual sum (Pr_{GPCC}) were applied to calculate Pr_M from Pr_{ann} (Eq. 10).

$$\mathrm{Pr}_{ann} = \mathrm{Pr} \cdot \mathrm{mean}\left\{\mathrm{Pr}'_{ann} / \mathrm{Pr}'\right\} \qquad \text{Eq. 9}$$

$$\mathrm{Pr}_{M} = \mathrm{Pr}_{ann} \cdot \mathrm{mean}\left\{\mathrm{Pr}_{M,GPCC} / \mathrm{Pr}_{GPCC}\right\} \qquad \text{Eq. 10}$$

We regressed annual values of critical SY to annual values of the climatic and hydrological factors to explain their inter-annual variability. The regression analysis not only aimed at finding the highest explained variability for all monitoring gauges but also to test the regression models for a more general applicability. To achieve this, the regression model for each gauge was compared to the regression model for all other gauges. In each case, we used t-tests to test the null hypothesis H0 that the exponents of both regression models do not differ significantly. The fewer the rejections of H0 were, the more suitable for a broader application the global regression model was considered.

3.5 Results and discussion

3.5.1 Evaluation of validation data

3.5.1.1 Interpolation of daily average SS concentration

The interpolation of daily data from irregular monitoring was the basis for the estimation of (critical) SS yields and model evaluation. Sampling frequency is always a trade-off between an adequate description of the natural SS variability and the cost for sampling and analysis. However, large deviations between both interpolation algorithms occurred during measurement gaps close to events with high SY and low SSC (Fig. 9) or inversely.

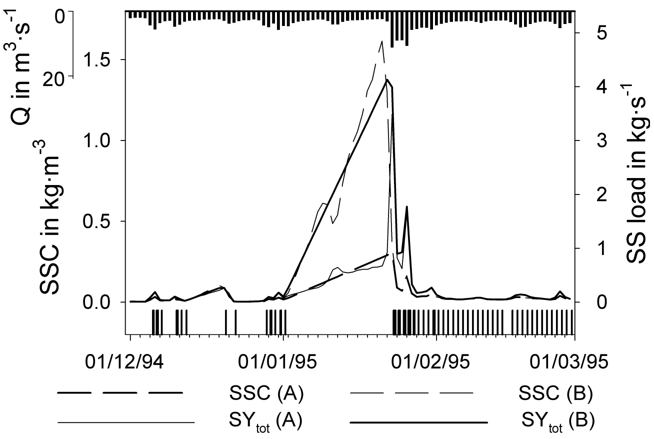

Fig. 9: Interpolation method and sampling interval (vertical lines at the bottom), gauge Hohenkammer

For model evaluation, the algorithm choice of the data provider had a significant ($p<0.0001$) though variable impact on interpolated SSC. Long-term and monthly means revealed 7% higher SSC values for interpolation method B than A with individual differences ranging from 1% to >20%. Significant and almost identical deviations for resultant long-term means were found for SY_{tot}, SY_{graph}, and SY_{FSD}. However, annual concentrations and yields were highly correlated ($r_S \geq 0.9$). Thus, the alternatives did not differ qualitatively and the spatial and temporal patterns for the model evaluation were not affected.

3.5.1.2 Calculation of critical SS yields

Long-term average critical SS yields were well correlated (r_S=0.87 for all, r_S=0.76 for non-alpine catchments). SY_{FSD} were on average about 25% lower than SY_{graph}, but individual ratios ranged from 0.4 to 1.8. SY_{FSD} was exceptionally overestimated for Taferlruck, a completely forested catchment where erosion events due to surface runoff were assumed to be rare. In most years water discharge was low, and occasional increases of SS concentrations and yields were merely due to base load variations, rather than soil loss in the catchment.

These findings show that critical SY were in general consistent though method choice affected model evaluation for single gauges and single years. Correlation coefficients for annual values were usually significant with few exceptions (Tab. 14). The estimation of critical SY affected the model evaluation more than the interpolation algorithm. Modifying the calculation of SY_{graph} had mostly minor effects on long-term averages. The annually calculated SY_{ann_graph}, however, converged to SY_{FSD} with respect to the inter-annual variability and the problematic long-term averages for Taferlruck and Kalteneck, further downstream.

Tab. 14: SY_{graph} alternatives compared to SY_{graph} and SY_{FSD}, correlation coefficients r_S were calculated with annual values

Gauge	Long-term average critical SY, $Mg \cdot km^{-2} \cdot a^{-1}$					r_S between SY_{FSD} and	
	SY_{graph}	SY_{FSD}	SY_{graph} alternatives			SY_{graph}	SY_{ann_graph}
			SY_{alt_bin}	SY_{tot_base}	SY_{ann_graph}		
Burghausen	110	64	98	101	134	0.87	0.87
Dietldorf	4.9	2.2	4.4	4.0	4.7	0.82	0.87
Gemünden	6.7	3.0	6.3	5.3	5.6	0.49	0.65
Kalteneck	5.8	10.6	6.3	5.5	8.9	0.80	0.68
Mainleus	3.9	3.8	3.4	2.5	4.3	0.46	0.83
Offingen	5.5	5.1	5.4	5.3	6.9	0.91	0.92
Plattling	5.1	3.7	4.8	4.4	7.2	0.77	0.83
Ruhstorf	11.7	11.9	11.7	10.4	10.4	0.86	0.85
Schlehenmühle	37	26	32	32	34	0.85	0.94
Taferlruck	0.5	4.6	0.5	0.4	2.4	0.61	0.37

3.5.2 Evaluation of model results

3.5.2.1 Soil erosion

We considered our catchment-wide soil losses to be consistent with those used by Auerswald (1992) and Behrendt et al. (1999). Applying their SDR models to soil loss and comparing the modelled SY to total SY (Auerswald 1992) and SY_{graph} (Behrendt et al. 1999) gave slopes of linear regression models close to 1.0. Although input data, temporal coverage, and the calculation of USLE factors slightly differed, r^2 values of 0.62 and 0.91 were also similar to, or even above, those published. Deviations were systematic. For instance, the underestimation of about 15% when applying Auerswald's approach occurred due to the USLE P factor of 0.85 for arable land in our study instead of 1.0. The soil losses were not evaluated for four catchments for which we assumed that critical SY values did not match the erosive fraction of total SY (next section).

3.5.2.2 Sediment yields and sediment delivery ratios

Correlations between area-specific soil erosion and critical SY were suspiciously weak for non-alpine catchments (cf. Tab. 17). Four gauges played an important role but no common explanation could be found. Most plausible reasons are SS sources outside the scope of the USLE that both disaggregation methods were unable to separate (urban runoff, in-stream erosion), sediment retention (not included in the SDR approach) and one very short, not representative time series. Without these gauges, correlation coefficients increased to $r_S \geq 0.7$ for SY_{graph}, but remained lower for SY_{FSD} ($0.5 \leq r_S < 0.7$).

Tab. 15: Regression models between alternative modelled and critical SY obtained for the 100m-DEM, model $y = a_1x$ with modelled SY as y and critical SY as x, t-tests used to test for the significance of $a_1 \neq 1.0$

Critical SY		Modelled SY				
		β_{Nbh}		β_{max}		β_{Nbh}
		L_{GIS}	L_{100m}	L_{GIS}	L_{100m}	L_{emp}
$SY_{A, graph}$	a_1	1.49**	1.40**	1.09	1.08	1.47***
	$r^{2\,(b)}$	0.59	0.59	0.61	0.64	0.69
$SY_{B, graph}$	a_1	1.28*	1.21	0.94	0.94	1.30**
	$r^{2\,(b)}$	0.51	0.52	0.51	0.56	0.65
SY^{a}_{FSD}	a_1	1.40*	1.33*	1.02	1.05	1.50**
	$r^{2\,(b)}$	0.39	0.37	0.37	0.40	0.56

[a] Without gauges at Taferlruck, Dietldorf, Bad Kissingen (see text for details)
[b] For intercept ≠ 0, *** p<0.001, ** p<0.01, * p<0.05

The comparison of critical SY to modelled SY revealed that model results deviated on average between 2 and 50% depending on the slope algorithm (Tab. 15). Although the lowest slopes of the regression models differed insignificantly from 1.0, all model variants could only moderately explain the observed spatial variability of long-term average critical SY. The repeatedly highest r_S and best explained variability showed that L_{emp} is the best description for erosive slope lengths in the study area. The resolutions of DEM and land-use data proved to be too coarse to derive suitable L_{GIS}. Deviations to SY_{graph} were largely the consequence of the underlying β_{Nbh}, and would diminish when replaced by β_{max}. The lack of data for one gauge was the reason for the higher r^2 for interpolation method A than method B.

Despite the good correlation with SY_{graph}, r^2 values dropped considerably when SY_{FSD} values were compared to model results ($0.14<r^2<0.29$). For two of the three most influential gauges, the modelling approach overestimated the critical SY. For the third gauge, Taferlruck, the SY_{FSD} estimation was problematic (section 3.5.1.2). However, r^2 remained lower even when these gauges were excluded.

The β_{max}-based SY_{graph} differed insignificantly from the modelled SY_{graph} of Behrendt et al. (1999). The average difference lay within the interpolation method's domain of uncertainty. The agreement to Behrendt et al. (1999) suggests that their large calibration factor of 0.25 can be noticeably reduced when considering flow distance.

The low, and even negative, correlation coefficients between modelled SDR and SDR derived from critical SY ($r_S<0.14$) suggested that the SDR equation is incomplete. This also holds true for the SDR equations of Auerswald (1992) and Behrendt et al. (1999). We thus conclude that the SDR approach needs to be adjusted (section 3.6).

3.5.3 Topographic uncertainty

In most cases, the choice of the topographic algorithm significantly influenced estimations of specific soil loss and sediment input (Tab. 16), although the differences between the slope and L alternatives were generally small for soil loss. Even though L_{emp} and its alternatives differed considerably for alpine catchments ($\beta>20\%$), USLE-based approaches explain only small fractions of the erosion processes and sediment yields there.

This effect was also observed for modelled SY (SY_{mod}), indicating that the differing spatial pattern of L alternatives did not influence model results. The slope method, however, noticeably affected model outcomes, as spatial distribution results in higher local SDR_{Nbh} near streams (section 2.4.4). The effect of lower short-distance SDR (b_{100}=1.19, b_{50}=1.07) was enhanced by excluding more upstream areas when β_{max} was applied (b_{100}=1.35, b_{50}=1.14).

Tab. 16: Correlation coefficients r_S and linear regression models $y = a_1x$ for modelled soil loss and sediment yield calculated with topographic alternatives. *indicate the significance of differences (Wilcoxon tests)

Correlated pair of values		β_{Nbh} - β_{max}			L_{100m} - L_{GIS}		L_{emp} - L_{GIS}	50 m - 100 m (L_{GIS})	
Alternatives		L_{GIS} (50 m)	L_{GIS} (100 m)	L_{100m}	β_{Nbh}	β_{max}	β_{Nbh}	β_{Nbh}	β_{max}
Soil loss (E) $Mg{\cdot}km^{-2}{\cdot}a^{-1}$	p	**	***	***	***	-	- (***[a])	***	***
	r_S	1.00	1.00	1.00	0.99	0.99	0.91	0.99	0.99
	a_1	0.98	1.02	0.97	0.93	0.97	0.97	1.22	1.28
Sediment yield (SY_{mod}) $Mg{\cdot}km^{-2}{\cdot}a^{-1}$	p	***	***	***	***	-	* (-[a])	***	***
	r_S	1.00	0.99	0.99	0.99	0.99	0.96	0.95	0.95
	a_1	1.14	1.35	1.29	0.92	0.96	0.95	1.88	2.23

[a] β≤20% (non-alpine catchments), *** p<0.001, ** p<0.01, * p<0.05, - not significant at the 5% level

Increasing the DEM resolution raised soil loss estimates by more than 20% (L_{GIS}). The minor effect of shorter erosive slope lengths and lower L factors (-8%) was dominated by higher slope angles. For L_{emp} and L_{100}, the dependency on DEM resolution was thus even more pronounced. The effects of DEM resolution on SY_{mod} were distinctly higher than on soil loss because of the denser drainage network in the 50m-DEM, resulting in shorter flow distances. Similar to chapter 2, the high correlation coefficients showed that topographic choices are of minor relevance for model quality except for the L factor (sections 2.4 and 3.5.3). This result also agrees with other observations (Wu et al. 2005; de Vente et al. 2009).

3.6 Improving model results

3.6.1 Controlling factors of spatial variability

The correlation analysis of the whole dataset revealed that SSC and q_{fast} almost equally well explained the distribution of long-term average SY (Tab. 17). Topography superimposes the relationship between catchment area A and SY_{tot} observed in many other studies. Steep slopes, high precipitation, and a large fraction of open spaces make alpine catchments effective sediment suppliers, independent of their area. Consequently, the correlations were strong for SY_{tot} (r_S>0.8) and catchment properties like the proportion of open space, average height (H), and precipitation (Pr) when assessing the whole dataset. This also held true for the SDR because the USLE underestimates soil loss in such catchments. Due to the high correlations, method choices had minor effects on correlation coefficients (except for the USLE L factor).

Tab. 17: Correlation coefficients between alternative SY, calculated SDR and catchment properties in Tab. 13 (neighbourhood slope method). Values in the first row were calculated with all catchments, second row without alpine, and third row without problematic catchments (section 3.5.2). Highest absolute values in bold font

	Land use	A	Morpho-metry	E	USLE factors w/o R[a]	q_{tot}	q_{fast}	Pr[b]	PP	SSC	Mixed w/o H·Pr[c]
$SY_{B, graph}$	-0.4 – 0.5	-0.0	-0.3 – 0.5	0.2	-0.3 – 0.6	0.5	0.6	0.6	0.4	**0.7**	-0.3 – 0.1
	-0.2 – 0.1	-0.2	-0.3 – 0.3	0.3	0.0 – 0.2	0.1	0.4	0.2	0.2	**0.6**	0.1 – 0.4
	-0.3 – 0.4	-0.2	-0.2 – 0.4	**0.7**	-0.1 – 0.6	0.1	0.2	0.2	0.4	**0.7**	0.4 – **0.7**
$SY_{B, FSD}$	-0.4 – 0.3	-0.2	-0.2 – 0.6	0.1	-0.4 – 0.6	0.6	**0.8**	**0.7**	0.3	**0.8**	-0.5 – 0.1
	-0.3 – 0.0	-0.4	-0.4 – **0.5**	0.0	0.0 – 0.3	0.2	**0.7**	0.4	0.0	**0.7**	-0.1 – 0.1
	-0.4 – 0.4	**-0.6**	-0.4 – **0.5**	**0.5**	-0.1 – **0.5**	0.1	**0.6**	0.3	0.2	**0.7**	0.3 – **0.5**
$SDR_{B, graph}$	-0.6 – 0.5	0.0	-0.4 – 0.6	-0.2	-0.6 – 0.5	0.6	0.6	**0.7**	0.3	0.6	-0.5 – -0.1
(L_{GIS})	-0.3 – 0.4	-0.1	-0.1 – 0.4	-0.3	-0.4 – 0.1	0.3	0.4	**0.5**	0.1	0.4	-0.3 – -0.2
	-0.2 – 0.1	0.0	-0.1 – **0.5**	0.1	-0.1 – 0.2	0.3	0.2	**0.5**	0.3	0.3	0.1 – 0.3
$SDR_{B, graph}$	**-0.7** – 0.5	0.0	-0.4 – 0.6	-0.6	-0.6 – 0.6	0.6	**0.7**	**0.7**	0.3	0.6	-0.6 – -0.1
(L_{emp})	-0.4 – 0.3	-0.1	-0.1 – 0.4	-0.3	-0.4 – 0.1	0.4	**0.5**	**0.5**	0.0	0.4	-0.3 – -0.1
	-0.2 – 0.0	-0.2	0.0 – 0.4	0.0	-0.1 – 0.1	0.5	0.4	**0.6**	0.2	0.2	-0.1 – 0.2
$SDR_{B, FSD}$	-0.6 – 0.5	-0.2	-0.2 – **0.7**	-0.4	**-0.7** – 0.5	0.6	**0.8**	**0.7**	0.1	0.6	-0.6 – -0.2
(L_{GIS})	-0.4 – 0.2	-0.4	-0.2 – **0.5**	**-0.6**	**-0.5** – 0.1	0.3	**0.7**	**0.5**	-0.2	0.4	-0.4 – -0.2
	-0.7 – 0.0	**-0.6**	-0.2 – **0.6**	-0.3	-0.3 – 0.0	0.2	**0.6**	**0.5**	-0.1	0.2	-0.1
$SDR_{B, FSD}$	**-0.7** – 0.4	-0.1	-0.3 – **0.7**	**-0.7**	**-0.7** – 0.6	**0.7**	**0.8**	**0.8**	0.2	0.6	**-0.7** – -0.1
(L_{emp})	-0.4 – 0.3	-0.3	-0.2 – **0.5**	**-0.5**	**-0.5** – 0.2	0.4	**0.8**	**0.5**	-0.1	0.4	-0.4 – -0.1
	-0.5 – 0.0	-0.4	-0.1 – 0.4	0.0	-0.1 – 0.1	0.2	**0.6**	**0.5**	0.2	0.4	0.0 – 0.2

[a] Identical to Pr, [b] MFI not shown here because was almost identical, [c] Close to Pr

For non-alpine catchments, the correlations between critical SY, SSC and q_{fast} were also pronounced, albeit less than for the whole dataset. The other catchment properties (Tab. 13) did not reflect the spatial variability of suspended solid yields in the study area, due to the wide range of environmental conditions and human activities. Topography (H, β_1) and topography-related properties (proportion of arable land and open spaces) are of less importance. The reasons for the low correlation between modelled soil erosion E and critical SY are discussed in section 3.5.2. Without the problematic catchments, correlations were considerably stronger. The strongest correlations to different SDR showed that the SDR model lacks a hydrological parameter, similar to those proposed by Auerswald (1992) and Behrendt et al. (1999).

3.6.2 Controlling factors of temporal variability

The water discharge (Q) better explained critical SY values than the annual precipitation (Pr), Modified Fournier Index (MFI) or suspended-solids concentration (SSC) (Fig. 10, Tab. 18). Although the estimated surface runoff Q_{fast} was the best parameter from which to predict critical SY, it is more feasible under most circumstances to use precipitation instead. This merely practical aspect was supported by t-tests that showed overall models to be more representative when Pr was the explanatory variable. The comparison of gauge-specific and general regression models revealed that both exponents differed significantly in only one case for Pr-based but considerably more often for Q-based and MFI-based models. Despite larger deviations, Pr is thus more suitable for regional applications than is Q_{fast}, while the lower residuals for Q_{fast} are advantageous when focusing on specific gauges.

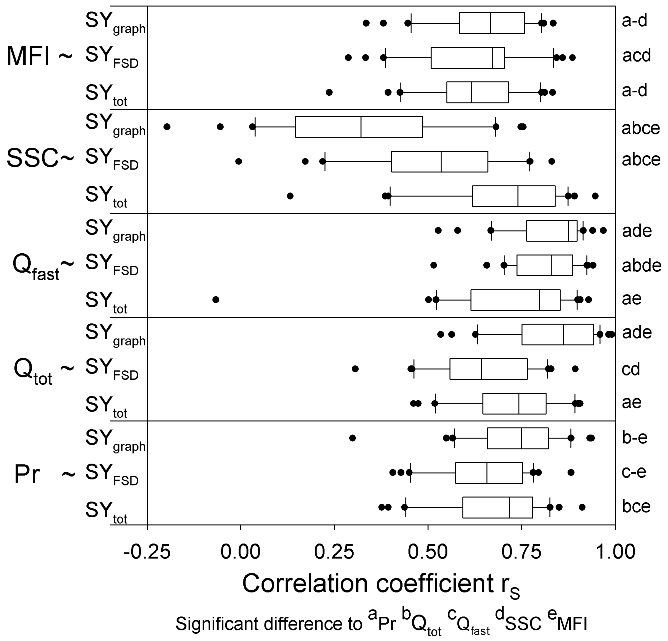

Fig. 10: Correlation coefficients for individual monitoring gauges between annual values of potential controlling factors and SS yields. Wilcoxon test performed for paired controlling factors. Significance shown if $p<0.05$

A visual analysis revealed that power equations were more suitable regression models than formerly proposed linear equations between hydrological parameters and SY (Auerswald 1992; Behrendt et al. 1999). Nonetheless, the generally high residuals (Tab. 18) show the limitations of aggregated values to explain the inter-annual variability of (critical) SY in this environmentally diverse study region. The highest exponents and residual standard deviations of regression models for SY_{graph} reveal the disproportionate scatter of annual values in comparison to SY_{tot} and SY_{FSD}.

It is important to note that all regression models clearly underestimated the SY values for a few station-years, mainly flood years in alpine catchments. Floods often occur during short time periods, thus more detailed input data is necessary to predict such extreme events. Additionally, such simple regression models cannot consider monitoring gaps during flood events and reservoir flushing (river Salzach in 1991) or sudden increases in SSC with constant Q due to works or collapsing banks (river Aisch, May 1983, M. Knott, pers. comm.).

Tab. 18: Regression models for critical annual SY compared to total SY for comparison (upper row for all catchments and lower row for non-alpine catchments). Power equations were derived by nonlinear least squares fitting

x	SY_{FSD}	SY_{graph}	SY_{tot}
Pr	$0.95 \cdot x^{2.89}$ (σ=0.56)	$0.84 \cdot x^{4.80}$ (σ=0.68)	$0.98 \cdot x^{2.19}$ (σ=0.37)
	$0.94 \cdot x^{2.91}$ (σ=0.52)	$0.81 \cdot x^{4.95}$ (σ=0.60)	$0.97 \cdot x^{2.19}$ (σ=0.34)
MFI	$0.97 \cdot x^{2.54}$ (σ=0.59)	$0.92 \cdot x^{3.74}$ (σ=0.76)	$0.99 \cdot x^{1.90}$ (σ=0.40)
	$0.96 \cdot x^{2.42}$ (σ=0.56)	$0.91 \cdot x^{3.69}$ (σ=0.73)	$0.98 \cdot x^{1.82}$ (σ=0.38)
Q_{fast}	$0.99 \cdot x^{1.26}$ (σ=0.46)	$0.95 \cdot x^{1.62}$ (σ=0.63)	-
	$0.99 \cdot x^{1.19}$ (σ=0.41)	$0.94 \cdot x^{1.60}$ (σ=0.57)	-
Q_{tot}	$0.98 \cdot x^{1.70}$ (σ=0.57)	$0.91 \cdot x^{2.78}$ (σ=0.61)	$0.99 \cdot x^{1.45}$ (σ=0.34)
	$0.97 \cdot x^{1.68}$ (σ=0.51)	$0.89 \cdot x^{2.79}$ (σ=0.52)	$0.99 \cdot x^{1.40}$ (σ=0.30)

3.6.3 Adjusting the sediment delivery ratio (SDR)

The moderately explained spatial variability proved the limitations of the modelling framework (Tab. 15). Irrespective of quantitative differences between the alternative model results, the high correlation coefficients demonstrated that, especially for topography, data and algorithm choices would not help to lessen this unexplained variability. An exception is the empirical approximation of the USLE L factor (L_{emp}) which proved to be superior to both alternative approaches. Nonetheless, strong correlations also showed that empirical, catchment-based models can efficiently cope with topographic uncertainty by including simple regression models. According to the correlation analyses (Tab. 17), a hydrological catchment property was included in the SDR model (Eqs. 11–12).

$$SDR_{cor,emp} = 0.011 \cdot SDR_{emp} \cdot q_{fast} + 0.95 \quad \left(r^2 = 0.46, SDR_{emp} < 10\%, n = 19\right) \qquad \text{Eq. 11}$$

$$SY_{cor,emp} = 0.90 \cdot SY_{B,graph} + 0.55 \quad \left(r^2 = 0.78\right) \qquad \text{Eq. 12}$$

The adjusted SDR values for non-alpine catchments were between 2.5 and 6.8%. The inverse correlation to catchment area (r_S=-0.24) was stronger than for the original SDR (Fig. 11). The explained variability increased significantly even though one gauge was included which was excluded before. The average deviation for the L_{emp}-based SY estimates decreased from +30% (Tab. 15) to -3%. This lay within the degree of uncertainty due to the interpolation methods for SS data. For large-scale applications, q_{fast} can be replaced by Pr. Results similarly improved for SY_{FSD}. The r^2 value rose significantly even with the inclusion of three gauges not previously considered (r^2=0.64) and average deviation was also -3%.

The remaining unexplained variability and the model residuals have to be seen in the model scope which did not include the numerous anthropogenic activities, long-term deposition and other discharge-dependent sediment sources. In accordance with findings of Wu et al. (2005), the inherent limitations of the empirical modelling framework may not allow for much more accuracy.

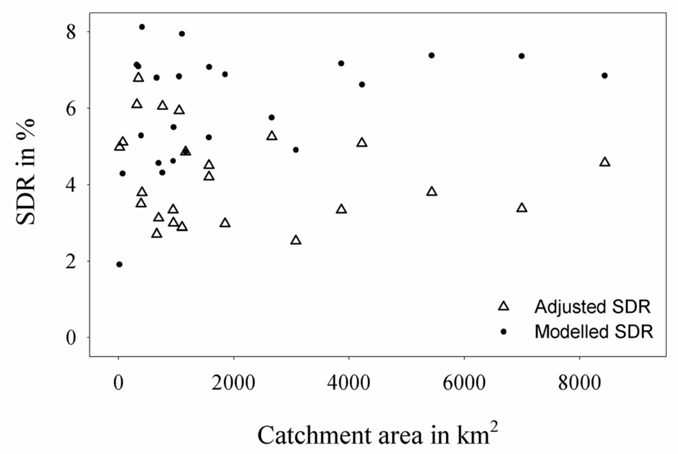

Fig. 11: Adjusted SDR and modelled SDR for non-alpine catchments (SDR modelled with L_{emp} as USLE L factor and SY_{graph})

The hydrological catchment property used for model improvement can be considered as proxy for catchment-wide hydraulic connectivity. Precipitation and surface runoff are not only driving forces of soil mobilization but also of sediment transport. With this adjustment, catchments receiving more precipitation or having more surface runoff are not only prone to more soil loss but to a higher proportion of sediment leaving the catchment as well. However, we are aware that the need for the hydrological adjustment also refers to our estimation of the USLE R factor, despite the successful verification of the modelled soil erosion. This concerns the interpolation of long-term average precipitation as well as the suitability of the linear empirical equations to estimate the R factor from long-term annual precipitation.

3.7 Conclusions

We have demonstrated that established empirical approaches for soil erosion and sediment delivery, in combination with Europe-wide available input data, can sufficiently explain the spatial pattern of long-term average yields of the erosion-related (critical) fractions of total suspended solid yields in hilly and mountainous catchments in southern Germany. To achieve a better explained variability of sediment yields in the study area, it was necessary to include a hydrological catchment property in the SDR approach. We expect that this modelling framework can also be applied in other European regions. Land use, flow distance (drainage density), hydrology (precipitation), and topography are commonly available for river catchments and thus suitable to estimate catchment-wide sediment delivery ratios in large-scale studies. However, within such modelling framework, multiple algorithms and DEM are available for derivation of topographic parameters and creation of validation data. Each alternative has a significant quantitative influence on topographic parameters, model estimates and validation data, respectively.

Our findings revealed that among these choices, DEM resolution had the largest impact on modelled soil erosion and sediment yields, with scale effects on slope angles and drainage density being most important. Model estimates were also strongly related to the choice of the slope algorithm with results differing by 30% between the neighbourhood and the maximum slope method. This value was close to the average deviation of 25% between both approaches to obtain critical fractions of total yields for model validation.

We also found that the estimation of critical yields affected the prediction of the inter-annual variability of sediment yields. Unlike the spatial variability, the mathematically-founded FSD approach gave better results than the alternative graphical approach. However, our aggregated hydrological parameters only moderately explained the inter-annual variability. Although we used power equations instead of formerly proposed linear regression models, extremely high sediment yields were still clearly underestimated. More detailed input data are needed to improve predictions for flood years.

The inherent limitations of the empirical modelling framework, and coarse data resolution, may impede determination of which alternative algorithm is superior. The range of relationships between model outcomes and validation data has to be seen as methodical uncertainty. Nonetheless, high correlation coefficients of long-term averages and annual critical suspended solid yields in most cases showed that the tested algorithms and DEM resolutions alone have only minor impacts on the explained variability of critical yields.

The erosion estimates were consistent with other studies. However, the calculated critical SDR were uncorrelated to our and other modelled SDR. Correlation analyses between critical SDR and catchment properties revealed that empirical SDR models should consider hydrology (surface runoff, precipitation). Consequently, including our estimated surface runoff significantly improved the model evaluation.

Although this study discussed the consequences of method choices for an USLE context in a complex region, the identified principles are also relevant for other modelling frameworks and regions. First, the estimation of total and subsequently critical sediment yields will affect any model calibration and validation. However, our observed similarity is limited to different interpolation approaches and may not be valid for different measurement methods and sampling frequencies. Second, the qualitative similarity of even complex topographic parameters suggests that catchment-based models can use regression models for any region to handle topographic uncertainty. Third, calibrating erosion models to total suspended solids overestimates the contribution of soil erosion. We recommend using critical yields instead and applied two approaches to estimate such critical sediment yields from daily monitoring data. In general, the results were consistent and methodical modifications were of minor relevance. However, the data-sparing and easy-to-use FSD approach agreed less to the spatial pattern of modelled long-term sediment yields but described the inter-annual variability of critical yields much more reasonably than the alternative graphical-statistical approach.

From our findings, we propose three topics for future studies. First, an improved estimation of the USLE R factor would make the SDR adjustment more reliable. Second, some ambiguous results and gauge-specific problems suggest that both approaches to derive erosion-related fractions from suspended-solids data should be further scrutinized. Third, the very high sediment yields of flood years have to be better captured.

4 Modelling the inter-annual variability of sediment yields: A case study for the upper Lech River

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4.1 Abstract

Pulsing storms and prolonged rainfall events have been associated to floods, soil erosion and nutrient fluxes in many European river catchments. This motivated us to develop a parsimonious approach to model the climate forcing on sediment yields in a mountainous Austrian-German river catchment. The hydro-climatologic forcing was interpreted by the novel RAMSES (RAinfall Model for SEdiment yield Simulation) approach to estimate the annual sediment yields. We used annual data on suspended-solid yields at the gauge Füssen, monitored from 1924 to 2003, and monthly rainfall data. The dataset was split into the period 1924–1969 for calibration and the period 1970–2003 for validation. The quality of sediment yield data was critically examined, and a few outlying years were identified and removed from further analyses. These outliers revealed that our model underestimates exceptionally high sediment yields in years of severe flood events. For all other years, the RAMSES performed well against the calibration set, with a correlation coefficient (r) equal to 0.83 and a Nash-Sutcliffe model efficiency (ME) of 0.69. The lower performance in the validation period (r=0.61, ME=0.36) has to be partly attributed to discontinuities in the monitoring strategy. For the calibration dataset, monthly precipitations proved nonetheless to be better predictors for annual sediment yields than annual values. These first results lay the foundation for reconstructing intra- to inter-decadal variability of sediment yields in river catchments where detailed precipitation records are not available as well as for the reconstruction of historical sediment yields.

4.2 Introduction

Understanding how climate forcing affects basin-wide responses is crucial not only for erosion and sediment modelling but also for the reconstruction of hydro-geomorphological hazards (Higgitt 2001; Angel et al. 2005). Climate and flow regime played key roles in the mobilisation and relocation of sediments in river basins (Walling and Fang 2003; Piégay et al. 2004; Macklin et al. 2006; Verstraeten et al. 2009). This was already documented in historical times, as scientists understood the erosion phenomena by empirical observations. For instance, Leonardo Da Vinci wrote in his "Treatise on Water" (1489): "The sediment of the rivers are more abundant in areas that are densely populated ... where the places, the mountains and hillslopes are tilled and the rain sweeps away more easily a ploughed soil than a soil that is hard and covered by vegetation" (translation by Pfister et al. 2009, p. 14).

Sediment yield is the amount of eroded soil exported over given time and space units. The underlying processes of soil erosion, sediment transport and deposition have been well described (e.g., Morgan 1986; Rose

1993; Haan et al. 1994) and many methods have been designed to assess soil erosion and sediment transport and the variability at different spatial and temporal scales (as reviewed by Brazier (2004) and Gao (2008)).

The many factors which control soil detachment and transport make sediment yields highly variable in space and time. Many empirical relationships have been proposed to relate the spatial variability to catchment properties (e.g., de Vente et al. 2007) because explicit modelling is demanding and not necessarily superior to empirical approaches (Jetten et al. 2003). Hydraulic connectivity and features such as channel distance, land use, and topography have been used to spatially disaggregate the sediment delivery of river catchments (e.g., Lenhart et al. 2005; Dymond et al. 2010; Fu et al. 2010).

The temporal variability is bound to rainfall power and overland flow, which trigger soil detachment and transportation. In many cases, a few large events contribute significantly to multi-year sediment yields (González-Hidalgo et al. 2010). Against the background of climate change, shifts in rainfall intensity and storm events are expected to affect future soil erosion and sediment yields (Wei et al. 2009). Annual sediment yields are usually weakly related to annual precipitation (Wilson 1973; Auerswald 1992). The inter-annual variability has been attributed to water discharge and climate pattern (Restrepo and Kjerfve 2000), the annual rainfall-runoff erosivity factor R of the USLE (Auerswald 1992), and heavy rainfall events (Lamoureux 2000). Welsh et al. (2009) even used a cellular automata model to simulate changes in the hydrological and sediment regime over 180 years at an hourly time step. However, the access to long records of water discharge and (sub-)daily input rainfall data for modelling is critical, and such data are only available for a few, geographically-scattered stations (e.g., Diodato and Bellocchi 2010).

This calls for parsimonious approaches for river catchments or time spans where the rainfall data are limited. Several simple empirical relationships between the R factor and commonly available factors have already been established (as, for instance, in Diodato and Bellocchi (2010)), mostly annual precipitation or simple indices based on monthly values. However, these factors may not sufficiently reflect the seasonal pattern of sediment processes like snowmelt in spring and thunderstorms in summer. Therefore, our aim was to develop and test an approach which i) explains the inter-annual and inter-decadal variability of sediment yields with a few, simple climate and land-use factors and ii) recognises seasonal and monthly precipitation associated with different sediment processes. Such an approach is necessarily unable to capture the details of individual sediment-forming fluxes.

We hypothesised that the monthly rainfall pattern is a better predictor for annual sediment yields than annual hydrological values. The novel model called RAMSES (RAinfall Modelling for SEdiment yield Simulation) was applied in the upper Lech River Catchment (LRC) to reconstruct annual suspended-solid yields. This catchment was chosen for two reasons. First, sediment data are available for 80 years, long enough to capture inter-annual to inter-decadal variations we are interested in here. Second, the upper Lech River has not substantially been affected by impoundment and regulation and allows studying the response of river catchments to recent environmental changes (Walling and Webb 1996).

4.3 Environmental setting and modelling

4.3.1 Study area

The study area was the catchment area of river Lech at Füssen, Germany (47° 34′ North, 10° 42′ East; Fig. 12). The Lech River is a major tributary in the upper Danube basin. The mountainous catchment covers 1,422 km^2 and is mainly placed in western Austria, with a small part in southern Germany. The average elevation is 1,615 m a.s.l. and the average slope is 46% as approximated from the CCM2 dataset (Vogt et al. 2007). On average, the annual precipitation total is 1,525 mm (data from HISTALP (Efthymiadis et al. 2006; ZAMG 2009)). According to our data, the long-term average suspended-solids yield is 235 $Mg{\cdot}km^{-2}{\cdot}a^{-1}$, with a distinct variability (interquartile range = 137 $Mg{\cdot}km^{-2}{\cdot}a^{-1}$). Land use is dominated by forests and alpine meadows (>60%), while arable land is sparse. Although the unfavourable environmental setting caused the decline of the agricultural sector during the last decades (Lantschner-Wolf 1990), the overall land cover did not change much since the 19th century (K. K. Statistische Central-Commission 1864, 1883; Statistik Austria 2010). For more details about the LRC, the reader may refer to Böhm and Wetzel (2006) and Engelsing (1988).

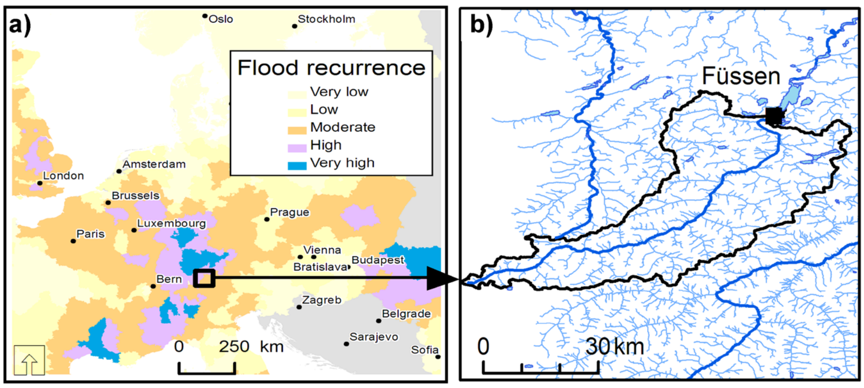


Fig. 12: Setting of the study area. a) flood recurrence classes (Schmidt-Thomé et al. 2006, modified), b) stream network as derived from CCM2 dataset (Vogt et al. 2007)

The climate is sub-continental with a pronounced intra-annual variability: 39% of the annual precipitation, 44% of the runoff and above 64% of the annual sediment yield occur between June and August (Engelsing

1988). For May to August, these values increase to 50–80%. The yearly and seasonal variability is the result of snow melt and of the synoptic circulation that advects air masses of different origins. This dynamic effect is strongly modulated by the Alps (Holton 2004). Air masses allow high storage of humidity that represents the main energy supply for thunderstorms, which are triggered by outbreaks of Atlantic polar maritime air in the middle troposphere.

Floods have been common in the LRC (Böhm and Wetzel 2006) (Fig. 12a) and have sometimes been severe (e.g., in 1910, 1999, 2002). Scheurmann and Karl (1990) provide a brief overview of the many flood protection measurements that were carried out in the upper Lech catchment. Nonetheless, the river partly maintained its natural wildness (Scheurmann and Karl 1990; Hettrich and Ruff 2011). In this region, the most extreme storm events are induced by instabilities produced by the frontal air masses brought from the North Atlantic, which result in both summer and early autumn precipitation maxima (e.g., Heino et al. 1999). This circulation pattern can be associated both with prolonged rainfall-driven large floods and sediment yield and intensive rainstorm-driven flash floods and intermediate floods with accelerated soil erosion (e.g., Diodato 2006). Large floods are accompanied by continuous runoff with great transport capacity of sediment at the basin outlet. Intermediate floods are suitable to remove great masses of soil that do not entirely arrive at the outlet, while net erosion driven by flash floods is negligible. In such Alpine catchments, snow melt can further increase runoff and sediment yield. In the LRC, the severe floods in 1910 and 1999 were the consequence of snowmelt and heavy rainfall (Meier 2004).

4.3.2 Data source

Annual values of runoff and total suspended solids were provided by the Bavarian Environment Agency for the years 1924–2003. They are similar to those discussed by Walling (1997). We used the HISTALP dataset to obtain monthly values of precipitation (Efthymiadis et al. 2006; ZAMG 2009). Information on land use was taken from the Corine Land Cover 2006 map (EEA 2010c) and statistical yearbooks.

Sediment data in Bavaria are measured as suspended solids. These are single-point measurements, which are occasionally calibrated and validated with multi-point measurements. The sampling frequency depends on water discharge. Samples are taken several times a day during runoff events or otherwise irregularly (daily to weekly). The average daily sediment yield is calculated from linearly interpolated daily suspended sediment concentrations and daily water discharge. More recently, the algorithm was changed to interpolating sediment loads directly, with considerable uncertainties when measurement gaps occurred next to flood events (section 3.5.1.1). Gao (2008) and Rode and Suhr (2007) describe general uncertainties associated with (suspended) sediment monitoring.

A few studies reveal 20–25% of the total sediment yield is transported as bedload (Maniak 2005; Turowski et al. 2010). Unfortunately, annual values are not available for long periods, so we could not consider bedload in our analysis of inter-annual variability. We implicitly assumed that bedload and suspended load changed similarly.

4.3.3 RAMSES model development

The nonlinear dependence of rainfall-runoff erosivity on rain intensity (e.g., D'Odorico et al. 2001) supports the adoption of nonlinear indicators for rainstorms. Actually, the use of nonlinear indicators as a measure of sediment yield was assumed to be a parsimonious approach able to be compared to the R factor of the USLE.

Although it is not possible to directly relate single heavy rainfall events to soil erosion types and, in turn, sediment yields, our hypothesis is that using predictors such as frequency and distribution of monthly rains can help defining modelling solutions for the varying sediment yield of rivers over long time spans. In this way, we used the same principle found in Diodato et al. (2012) to expand a satisfactory solution in which monthly rainfall aggressiveness and overland flows are modelled together to account for temporal dependence of rainfall-runoff. Runoff is generated by rainstorms and its occurrence and quantity depend on intensity, duration and distribution of rainfall events. The relationship between runoff and rainstorms may effectively be captured based on the intra-annual rainfall pattern. In RAMSES, this relationship is interpreted as a nonlinear response to monthly and seasonal precipitations (Eq. 13).

$$SY_{RAMSES} = \alpha \cdot \left(\sum_{Jun}^{Aug} Pr_M \right)^{\eta} + \nu \cdot \sqrt{\delta \cdot Pr_{Dec-1} + \sum_{Feb}^{May} Pr_M + Pr_{Sep} \cdot \max(Pr_{Aut}) - B} \qquad \text{Eq. 13}$$

SY_{RAMSES} is the estimate of the annual sediment yield (net erosion) in $Mg{\cdot}km^{-2}{\cdot}a^{-1}$. The first term represents the erosive component driven by the summer precipitation Pr (mostly as thunderstorms) between June (Jun) and August (Aug). In the second term (under the square root), three predictors represent the soil transport governed by the average rainfall-runoff during winter and spring (antecedent December (Dec-1), February (Feb) to May) and the peak runoff rate during autumn (Aut = September + October + November). The latter is a combination of a proxy of soil humidity (precipitation in September, Pr_{Sep}) and the monthly maximum rainfall in autumn (Pr_{Aut}). All rainfall inputs are expressed in mm. The term B is the long-term average gross erosion retained in the catchment.

The empirical parameters α, δ and η express the rainfall power within the climatic context. The coefficient ν is a vegetation exponential function describing the protection by land cover (Eq. 14, after Thornes (1990)). LVC is the land vegetation cover, and γ and ζ are two empirical parameters.

$$\nu = \gamma \cdot e^{-\zeta \cdot LVC} \qquad \text{Eq. 14}$$

4.3.4 Model evaluation

The dataset was split for model calibration (1924–1969) and model validation (1970–2003). Several statistical tests were applied to evaluate the agreement between RAMSES estimates and actual sediment values. First, the Pearson's correlation coefficient r and r^2 were calculated to assess the linear dependence between modelled and observed data and the variance explained by the model. Second, the Nash-Sutcliffe model efficiency (ME) (Nash and Sutcliffe 1970) and the mean absolute error (MAE) were derived to assess quantitative differences. For perfect models, MAE is 0 and ME is 1. Poor models have high MAE (up to +∞) and low ME (down to -∞). Third, the Durbin-Watson test (Durbin and Watson 1950, 1951) was conducted to test for autocorrelated residuals because strong temporal dependence may induce spurious correlations (Granger et al. 2001). All analyses were performed with STATGRAPHICS Online (StatPoint Technologies 2012).

We also compared the RAMSES to three simplified approaches (Eqs. 15–17). Pr is the annual precipitation (mm), MFI is the Modified Fournier index (FAO and UNEP 1977), and q is the total annual runoff ($l \cdot s^{-1} \cdot km^{-2}$).

Precipitation model $$SY_{PM} = a \cdot Pr^{b} \qquad \text{Eq. 15}$$

MFI model $$SY_{MFI} = a \cdot MFI^{b} \text{ with } MFI = \sum_{M=1}^{12} \frac{Pr_M^2}{Pr} \qquad \text{Eq. 16}$$

Runoff model $$SY_{q} = a \cdot q^{b} \qquad \text{Eq. 17}$$

These models have been developed in chapter 3 to describe the inter-annual variability of sediment yields in the same region between the 1970s and 2003. Their runoff model was slightly superior in explaining the variability of sediment yields, although catchment-specific regression models differed significantly. All models failed to predict the extraordinary high sediment yields of some flood years in alpine catchments (section 3.4.4.2).

4.4 Results and discussion

In this section, the comparison between measured sediment yield (SY) and RAMSES results is presented and discussed. The concept and the assumptions of RAMSES are discussed for its applicability beyond the study area.

4.4.1 Model parameterization and evaluation

Equations 13 and 14 were parameterized according to Tab. 19. A few observations were dropped from further analyses because of very high residuals (Fig. 13a). For the calibration dataset, we observed suspicious values for 1934, 1936, 1945, 1956, 1960 and 1965 but found no common explanation.

Tab. 19: Values assigned to parameters of Eqs. 13 and 14

Equation	Parameter	Value	Source
13	α	0.625	Calibration
	η	1	Calibration
	δ	50	Calibration
	B	235	Calibration
14	γ	170	Calibration
	ζ	0.07	Thornes (1990)
	LVC	81	Statistical yearbooks

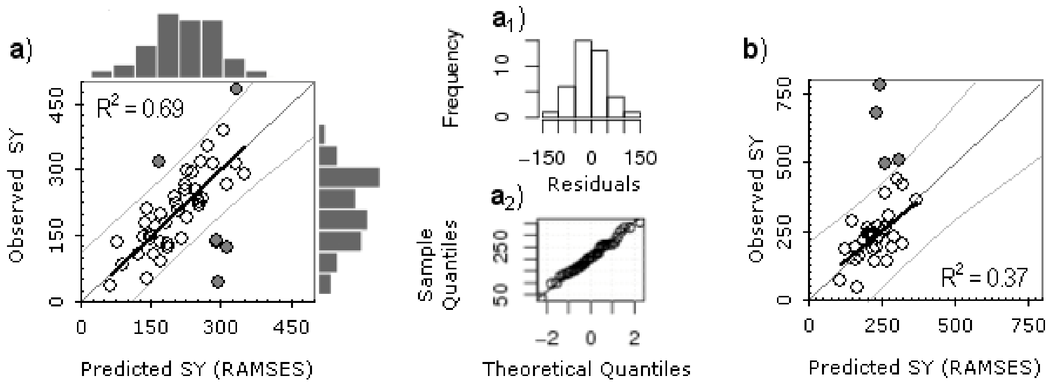


Fig. 13: Analysis for the calibration and validation datasets. a) scatterplot between predicted and observed sediment yields (Mg·km^{-2}·a^{-1}) for the calibration dataset with bounds of 95% prediction limits (grey curves), frequency distributions on both axes, 1:1 line (grey) and the line of best fit (black). The regression was calculated without the outliers (grey), a_1) distribution of residuals, a_2) comparison of theoretical to sample quantiles (QQ-plot), b) scatterplot for validation dataset similar to a)

We assumed that the sediment yield and the precipitation values for 1945 were affected by the (post-)war situation. In 1965, late snow-melt together with high rainfall led to a major flood in June (e.g., Embleton-Hamann 2007). For that year, RAMSES underestimated the very high sediment yield. For the four other outliers, the precipitation in June was above the third quartile of the whole dataset while winter precipitation was below average (Fig. 14). This indicates that the model in a few cases overemphasises the effect of the precipitation in early summer or during the winter on sediment yields.

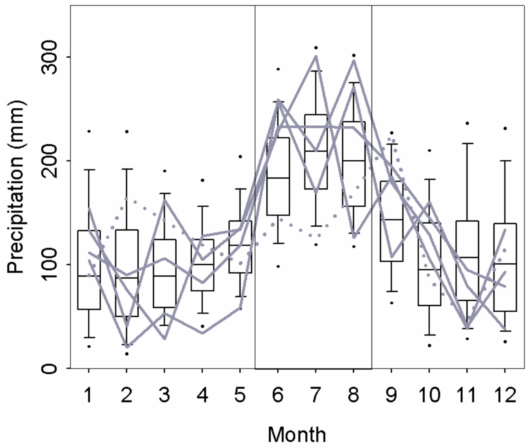


Fig. 14: Monthly precipitation regime for the Lech catchment (1924–2003) (box plots) compared to years with high residuals in the calibration period (dotted line for 1945, straight lines for all other years)

For the other 40 years, the explained variability and the model efficiency were satisfactory (Tab. 20). The Durbin–Watson test indicated that residuals were not autocorrelated.

Tab. 20: Performance and autocorrelation of the RAMSES model at calibration and validation stages

Dataset	Performance statistics			Autocorrelation statistics	
	Nash-Sutcliffe model efficiency	Correlation coefficient	Mean absolute error ($Mg \cdot km^{-2} \cdot a^{-1}$)	Lag-1 residual correlation	Durbin-Watson statistic (significance)
Calibration	0.69	0.83	38.0	-0.0738	1.718 (p=0.19)
Validation	0.36	0.61	53.4	-0.0881	2.145 (p=0.67)

At the calibration stage, the two main components of Eq. 13 were treated as independent variables, and each of them was tested for their significance on the output, as in a multiple linear regression. The highest p value belongs to the square-root term and is well below 0.01 (p<0.00001). The strength of both terms revealed that Eq. 13 should not be further simplified.

For model validation (Fig. 13b, Tab. 20), 4 years (1970, 1995, 1999 and 2002) affected strongly the model performance. Similar to 1965, severe floods occurred in these years and RAMSES clearly underestimated the associated high sediment yields. The model agreement was again satisfactory albeit lower than for the calibration period, even after excluding these outliers (Tab. 20). We assume that inconsistencies and the lack of homogeneity in the sediment records are relevant (e.g., Walling and Fang 2003). Over eight decades, the data have likely been obtained using different techniques (M. Knott, Bavarian Environment Agency, pers. comm.), and thus its quality is variable. Additionally, high-frequency sampling is particularly important during flood events, when often the major part of the annual sediment load is transferred to the outlet. Consequences of measurement gaps during such events observed at other Bavarian gauges are discussed in section 3.5.1.1. So, we have to assume that not all collected SY values at Füssen have been obtained under optimal conditions. Finally, the commercial removal of gravel and river engineering works affected the sediment transport during the last decades (Scheurmann and Karl 1990).

The comparison to the three alternative models revealed that RAMSES performed better because the residuals of each alternative model were larger and the explained variability significantly lower (Fig. 15). The performance of the alternative "runoff model" (r^2=0.67) is comparable to that of RAMSES (r^2=0.69). The former is indeed based on an input variable (total runoff) directly related to sediment yields, yet not readily available and therefore not usable for long-term estimations.

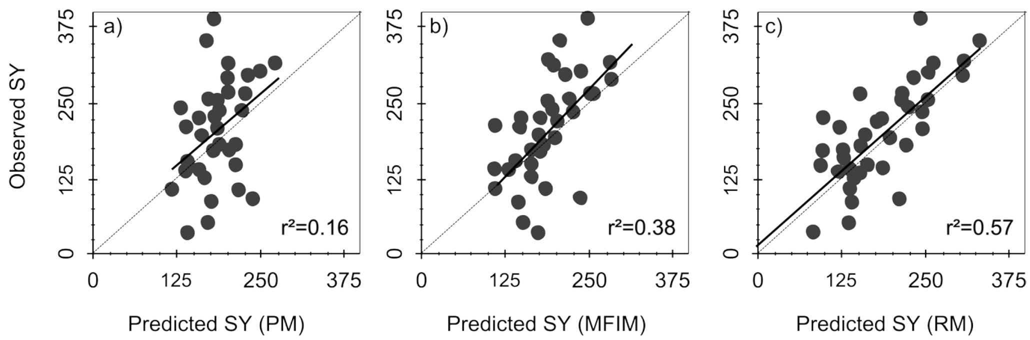


Fig. 15: Scatterplots between observed sediment yields and the alternative models ($Mg\cdot km^{-2}\cdot a^{-1}$). a) annual precipitation model ($0.0031\cdot Pr^{1.50}$), b) Modified Fournier Index model ($0.0033\cdot MFI^{2.16}$), c) runoff model ($0.1358\cdot q^{1.90}$). Coefficients were calibrated against sediment yield data for Bavarian gauges. Precipitation data was similar to the RAMSES calibration.

4.4.2 Discussion of RAMSES

4.4.2.1 Model structure

In streamflow modelling, hydro-climatological processes that are commonly quantified are precipitation, interception, evaporation, transpiration, snow accumulation and melt, soil water movement, and overland and channel flows (e.g., Koivusalo and Kokkonen 2003). The importance of each process depends on the catchment and climatic characteristics, as well as on the modelling objectives. In relatively small and steep mountain catchments, hydro-geomorphological processes are strongly characterised by nonlinear interactions between climate forcing, land surface processes, and fluvial responses over different spatial and temporal scales (Wolman and Gerson 1978; Huss et al. 2008; Brardinoni et al. 2009; Schiefer et al. 2010). Therefore, climate and geomorphological processes are difficult to model and show high variability in such an environment (e.g., Calanca et al. 2006). The main idea of RAMSES is that three hydro-climatological factors are relevant for the annual sediment yields in river basins: rainfall-runoff erosivity in winter and summer and transport capacity. As they occur at different times during a year, they have to be weighed depending on the seasonal pattern of precipitation. The structure of Eq. 13 is essentially based on the knowledge of the European climatology. Key climatological aspects are interpreted by the RAMSES structure, which make Eq. 13 potentially suitable for sediment yield estimations outside the studied catchment.

The first term includes summer precipitation as an important driver for rainfall-runoff processes in Central and Western Europe (van Delden 2001; Twardosz 2007). Using reanalysis data, Romero et al. (2007) described severe convective storms occurring between June and October in Europe. Especially in the vicinity of the Alps, thunderstorms tend to be more numerous during the warm season, due to frontogenesis and cyclogenesis mechanisms associated with the advection of warm air masses (van Delden 2001).

The second term of Eq. 13 includes the precipitation pattern during the other seasons. Additionally, the square-root is a mechanism to either attenuate or enhance rainfall-runoff erosivity depending on site-specific climate conditions. Runoff generation mechanisms describe different processes in response to rainfall and snow melt (and subsequent water movements). During autumn, rain events present homogeneous features most of the time in Western Europe, especially during October (van Delden 2001). In the Mediterranean region of Europe, the rains and storms become more extended with the continuing of the autumn season, thus exhibiting relevant geomorphologic effectiveness towards more runoff than the rainfall erosivity fraction (Diodato 2006). The dynamics of snow accumulation and melting may drive runoff generation in snow-dominated regions. Holko et al. (2011) describes the role of snow in the hydrological cycle in mountainous areas of Central

Europe. Typically, most of the water storage accumulates in the snowpack during the winter season, where snow melt generally represents low and relatively constant discharges (Bartolini et al. 2009). High and variable discharges are typical for snow melt in spring, which may lead to large seasonal runoff (López-Moreno and García-Ruiz 2004). For our modelling purposes, January was left out because it did not significantly contribute to the inter-annual variability of sediment yields. During this month, European conditions are generally favourable for snow cover forming and persistence (Bednorz 2009) without contributing to sediment yields.

In the region of the study area, intense rainfalls generally occur between the end of spring and start of autumn (dark-coloured line in Fig. 16a), when both rain-splash and runoff peaks are present. In the autumn season, rain rates considerably decrease (dark-coloured line in Fig. 16a), resulting in the predominance of overland flow in runoff formation only (which are able to redistribute sediment across the drainage basin). The precipitation from December to May leads to a hydrological regime related to snowmelt, which differs from June–November. In particular, frequent flash floods facilitate overland flow in summer (dark-coloured box in Fig. 16b), while high discharges in autumn (light-coloured box in Fig. 16b) depend on rainfalls of long duration and low intensity (Merz and Blöschl 2003).

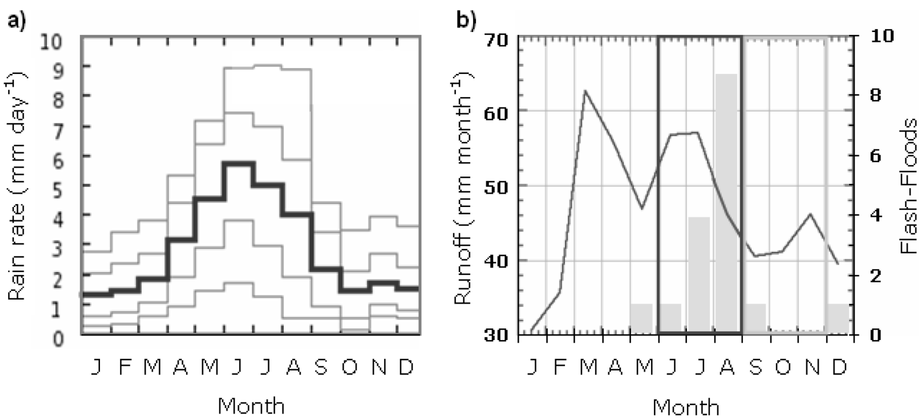


Fig. 16: Precipitation rates and hydrological regimes in area across the LRC. a) monthly regime of rain rate, b) monthly runoff (line) and frequency of flash floods (histogram). The rain rates were derived from reanalysis data from 1947 to 2007 (Kalnay et al. 1996). The dark-coloured line in a) is the mean, the light-coloured lines represent the 3rd, the 17th, the 83rd and the 98th percentiles (from the bottom to the top). The frequency distribution of flash floods was derived for Austria for 1947–2007 (Gaume et al. 2009; Gaume et al. 2010). The most powerful rain events in summer (dark-coloured box) and autumn (light-coloured box) are those expected to drive sediment yields.

The last component of the model is the sink term B. It represents the fraction of the gross erosion (E, $Mg \cdot km^{-2} \cdot a^{-1}$) retained in the catchment (Eq. 18).

$$B = (1 - SDR) \cdot E \qquad \text{Eq. 18}$$

SDR is the sediment delivery ratio, i.e. the ratio of sediment yield at the catchment outlet to total erosion in the catchment. The concept is an analogue to the connectivity ratio (the amount of sediment reaching a stream over the amount of sediment eroded), which characterises the efficiency of slope-channel transfer and depends on the transport capacity and slope-shape and drainage pattern (e.g., Quinton et al. 2006). In Alpine catchments, however, given the occurrence of landslides and other processes, B cannot be easily calculated. Rather, B is considered as a long-term constant depending on catchment characteristics like basin area, terrain and average annual precipitation (Diodato and Grauso 2009). B is not necessary to describe the inter-annual variability of sediment yields and may be omitted in applications where sediment yields are calculated relative to the long-term average.

Our purpose is to show that RAMSES adequately illustrates the importance of nonlinear interactions between climate forcing and landscape response that affect sediment supply. It also reproduces the behaviour of the observed mountain catchment because the optimised parameter values determined over a pool of years ensured a generic temporal representation of sediment yields. An important property of the model is that it does not reproduce well high-magnitude sediment transport events or years. These cases may reflect geomorphological thresholds and feedback effects, as described in other studies (Wolman and Gerson 1978; de Vente and Poesen 2005; Guzzetti et al. 2008). For instance, during periods of low-magnitude events, loose sediment accumulates on catchment slopes and will lead to a disproportionately large sediment supply during the next transport-triggering rainstorm. Conversely, if the sediment system is supply limited, only small amounts of sediments will be transported during events (Otto and Dikau 2004; Schlunegger et al. 2009; Schiefer et al. 2010). These characteristics of sediment dynamics in mountain catchments have been observed not only in the Alps, but also in Scandinavia (Beylich and Sandberg 2005; Bartsch et al. 2009) and the Canadian Rockies (Brardinoni et al. 2009; Schiefer et al. 2010).

4.4.2.2 Model calibration

In practical modelling studies, hydrologists have to cope with data scarcity. We are aware that our modelling study suffers from lack of data and that the RAMSES may be sensitive to variations in its parameter values. A priori, this has motivated the adoption of a differential split-sample approach. Rather than imposing arbitrary data-splitting schemes, the available data were divided according to a given criterion to show that the model concept is suitable to predict sediment yields for conditions different from those for which it was calibrated. The first set of old data (1924–1969) was much larger than the 20–30 years recommended to capture the natural sediment variability in river catchments (e.g., Lu et al. 2005). It also allowed us to capture the variability associated with data inhomogeneity, which is more likely in older records. This complies with recommendations by Sorooshian et al. (1983), who suggested having as much variability as possible in the calibration record, and is a common approach in other domains (e.g., Venåsa and Rinnan 2008). We used the smaller set of more recent data (1970–2003) to validate the model predictions. For the study catchment, this differential split-sample approach sounds appropriate to assert the reliability of the model for non-stationary conditions, which makes it useful to predict effects on sediment yields over long periods, with changing climatic conditions (Xu 1999a, b).

A posteriori, we have verified that the model parameterization obtained may assure acceptable performance (r^2>0.50) when used over either short (Fig. 17a–b) or extended (Fig. 17c) periods of time. In the three cases, the points tend to line up around the 1:1 identity line, with the three slopes (1.31, 0.90, 0.98 in a, b and c, respectively) not differing significantly (p=0.34).

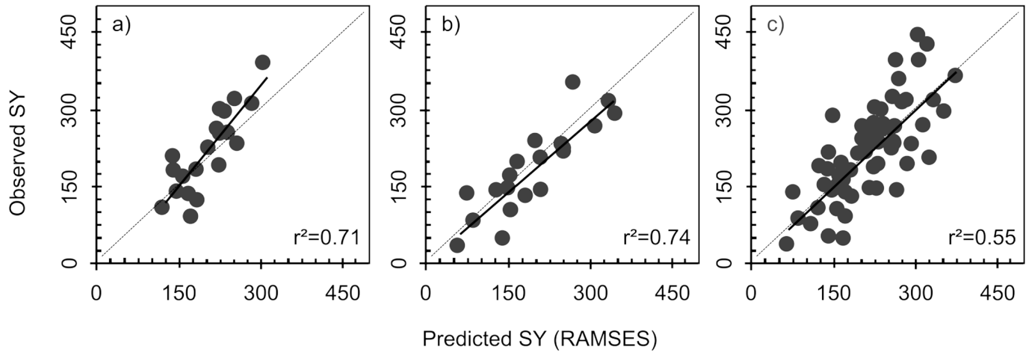


Fig. 17: Scatterplot between observed and predicted (RAMSES) sediment yields (SY, $Mg \cdot km^{-2} \cdot a^{-1}$) for different periods of time. a) 1924–1946, b) 1947–1969, c) 1924–2003

For future applications of the RAMSES model to other Alpine sites, it will be necessary to identify appropriate non-climatic factors like hydropower plants, reservoirs, or afforestation, which are likely to constrain the sediment system and may require local optimization.

4.5 Conclusions

Parsimonious hydro-climatological models are appealing for predicting sediment yields when high-resolution precipitation data is not available. However, the high inter-annual variability of sediment yields demands for a better understanding of the mechanisms responsible for it. A common problem is that the time series to develop or test models are quite short.

The novel RAMSES model was developed and successfully validated in a mountainous river catchment to predict annual sediment yields over eight decades from monthly rainfall data. The comparison to three alternative empirical relationships revealed that the prediction of the model was significantly better. The results suggest that a small number of parameters may be sufficient to represent annual sediment yields with enough detail. We also observed that the model is of limited use to capture extreme values occurring during severe floods. For these very few cases, a more detailed model is needed.

At the current stage of development, a few reasonable and physically sound assumptions serve to focus on the variability and evolving trends of sediment yields on the intra- to inter-decadal scale, and not on the inter-annual scale. The current RAMSES is a generic model, which lays the foundation for future applications in other river catchments and for the reconstruction of historical sediment yields. Although RAMSES was specifically developed for the study area, its general structure should be applicable to other European river catchments. Nonetheless, the model structure has to be refined in future studies. First, the significance of non-climatological factors other than land cover for the response of river basins to climate forcing has to be scrutinised. Second, the model has to be normalised to long-term averages to account for the range of natural characteristics in European river basins. Third, the seasonal windows assigned to different processes in RAMSES remain critical and have to be further validated.

5 Modelling soil erosion and empirical relationships for sediment delivery ratios of European river catchments

Submitted to *Catena*

5.1 Abstract

Sediment delivery ratios (SDR) are common tools to link soil loss to sediment yields (SY) of river catchments. Large-scale erosion models rely on simple catchment properties to estimate such SDR. In order to assess the sensitivity and uncertainty of these SDR models in different European regions, I compiled an extensive European sediment database and derived various soil loss maps from pan-European data. 16 maps were created using the universal soil loss equation (USLE) and different approximations for its R, L, C, and K factors. The sensitivity to USLE parameterisation was compared to the sensitivity to soil loss and SDR models. The SDR of river catchments were calculated from the USLE maps and a PESERA map and 4 regression models were applied to explain the variability of SDR and SY. As a consequence of the huge uncertainty in USLE estimates, the uncertainty in modelled SY ranged on average from 30% to 60% in different parts of Europe. The model and algorithm choices also affected the quality of SDR and SY predictions. Most relevant for the model efficiency and r^2 was the soil loss model. The combination of the physically based PESERA and SDR models could not explain the variability of SY. In contrast, using the USLE with the 4-parameter ("Complex") SDR model allowed satisfactory results in all regions. SDR models with fewer parameters were less appropriate. Among the USLE factors, the K factor had a stronger impact on the performance of the Complex model than the L and C factors. The sensitivity to the R factor was mostly low. Their combined impact on the model efficiency was as important as the parameterisation of the SDR model. Carefully choosing the soil loss map can thus significantly improve regional SDR and SY predictions. A new soil loss map is proposed as most suitable for predicting SY of European river catchments with the Complex SDR model. Nonetheless, no model realisation could grasp the huge variability of SDR and SY in Europe. Only regional applications were feasible. Large residuals were plausibly related to unrepresentative sediment data and limitations of the modelling framework.

5.2 Introduction

Soil erosion by water is one of the main threats to soil functionality, crop production, and water quality in Europe (Boardman and Poesen 2006a; EEA 2007b). Sediments and sediment-associated nutrients and contaminants negatively affect aquatic habitats (Bilotta and Brazier 2008), the lifetime of reservoirs (Einsele and Hinderer 1997), and intensify flood damage (Merz et al. 2010). These negative effects on our society and environment have been addressed by the European environmental policy and risk assessment (EC 2006; van Beek et al. 2010; Kibblewhite 2012).

Extensive studies have been conducted for decades and numerous approaches have been developed to quantify the mobilisation and relocation of soil particles as reviewed by de Vente and Poesen (2005) as well as Jetten and Favis-Mortlock (2006). Despite increasing computing power, the high variability of underlying processes and the demanding parameterisation limit the application of complex, spatially distributed models and do not necessarily make them outperform empirical models (Nearing 1998; Jetten et al. 2003; Govers 2011). For soil loss, the most common of such empirical models are based on the USLE (Eq. 1, p. 23) which considers topography (L·S), soil properties (K), land use and land management (C·P), and rainfall erosivity (R).

Despite conceptual flaws (Kinnell 2004), the USLE has still been widely used in Europe to estimate soil loss on regional (van der Knijff et al. 1999; Stumpf and Auerswald 2006; Bakker et al. 2008; de Vente et al. 2009; Kouli et al. 2009; Krása et al. 2010; Volk et al. 2010; Tetzlaff et al. 2013), country (Grimm et al. 2003; Rousseva et al. 2006a; Prasuhn et al. 2007; Strauss 2007; Cebecauer and Hofierka 2008), and continental scales (van der Knijff et al. 2000; Podmanicky et al. 2011). USLE concepts are also part of hydrological models like SWAT and nutrient models like MONERIS (Gassman et al. 2007; Venohr et al. 2011). Other pan-European estimations were derived with the physically based PESERA model (Pan-European Soil Erosion Risk Assessment) (Kirkby et al. 2008) and from measured soil loss rates (Cerdan et al. 2010). Van Rompaey et al. (2003b) provide a brief overview of earlier approaches.

Sediment yield (SY) is the quantity of eroded soil reaching the catchment outlet, i.e. the net erosion. The complex interaction of many factors and processes results in a high variability in space and time (e.g., Verstraeten et al. 2003; González-Hidalgo et al. 2009; Cantón et al. 2011; González-Hidalgo et al. 2013) and sophisticated models have been developed to predict SY accounting for sediment routing, deposition, and remobilisation (Lu et al. 2005). Physically based models use hydraulic variables to estimate changes in the transport capacity of surface runoff (Prosser and Rustomji 2000; Jetten and Favis-Mortlock 2006). However, the high data requirement limits these complex models to small areas. Empirical and conceptual models implement sediment delivery ratios (SDR) as adjustment factors to soil loss estimates instead.

The SDR of a catchment is defined as ratio of the (observed) SY at the outlet to the (modelled) soil loss. As soil erosion and sediment transport are highly variable in time, the SDR necessarily depends on the measurement period and reflects the time scale over which sources and sinks operate (Lu et al. 2005; Walling 2008). To obtain average SDR for regional studies, sufficiently long and consistent measurements of soil erosion and sediment yields are needed yet rarely available (Walling 1997; Lu et al. 2005).

To predict the SY and SDR of ungauged catchments, many empirical relationships between the SDR and catchment properties have been suggested, as reviewed by de Vente et al. (2007) and Walling (1983). The common power equations to predict SDR from the catchment area vary enormously between regions revealing a complex scale-dependency of soil erosion and sediment delivery. Explanatory factors such as geomorphology, climate, and land use have thus been proposed to improve the extrapolation of SDR (Roehl 1962; Walling 1983; Ichim 1990; Diodato and Grauso 2009; Venohr et al. 2011). Additionally, travel time, flow path properties, land use, topography and other features have been applied to spatially disaggregate SDR of catchments (sections 3.2 and 3.4.3.4).

By definition, any calculated SDR and any empirical SDR model depends on a soil loss map and sediment data. Both are subject to uncertainty due to algorithm and data choices among other sources. First, the SY can be measured as either fluvial transport or the sedimentation rate in reservoirs, lakes, or ponds and both methods give complementary results (Brazier 2004). The uncertainty in riverine sediment yield is related to the sampling frequency and how infrequent observations are interpolated and extrapolated (Webb et al. 1997; Phillips et al. 1999; Kauppila and Koskiaho 2003; Moatar et al. 2006) (section 3.5.1.1). For reservoir and pond data, the estimation of the trap efficiency and bulk density of sediments is crucial (Verstraeten and Poesen 2000, 2001; Verstraeten et al. 2003). Although fluvial sediments are typically transported in suspension, uncertainty also arises from the highly variable yet rarely measured bed load (Turowski et al. 2010). The numerous sources of uncertainty in determining SY make error estimations difficult. Vanmaercke et al. (2011) have concluded that errors may exceed 100% if the sampling is infrequent.

Second, the USLE and SDR parameterisation for large-scale applications is limited by the currently available input data which is insufficient to calculate USLE factors directly. Previous studies showed that the topographic parameters of the USLE are important sources of uncertainty (Renard and Ferreira 1993; Risse et al. 1993; Biesemans et al. 2000; Tetzlaff et al. 2013). For a single catchment, however, the uncertainty in soil loss is compensated by the SDR. While purely quantitative impacts of data and algorithm choices are thus of less relevance, changes in the spatial pattern indeed affect the calibration and application of empirical SDR models.

Uncertainty and sensitivity analyses at large scales have focused on hydrological and topographic parameters (chapters 2 and 3). They revealed strong topographic similarities, despite significant quantitative impacts of data and algorithm choices. In contrast, this study addresses the uncertainty in USLE factors and the consequences of soil loss estimation for the application of empirical SDR models in Europe. How uncertain are the modelled soil loss and SY? How sensitive is the explained variability to the soil loss mapping? Does one most suitable (but not necessarily more precise) map exist to satisfactorily predict the variability of SDR and SY of European river catchments?

5.3 Methods

5.3.1 Outline

The empirical modelling framework comprises the approximation of USLE factors from pan-European input data and the application of regression models to explain the spatial variability of SDR and SY of European catchments. The European context is important because the variety of soil loss rates, SY and SDR is huge in Europe. It was hypothesised that regional findings cannot be extrapolated. To obtain homogeneous model parameters and soil loss maps, pan-European input data was used. As the study aimed at uncertainty and sensitivity analyses rather than at a conceptually new soil loss map, previously published approximations of USLE factors were used for mapping the gross soil loss. These approximations were independently varied for each factor. To limit the number of soil loss maps, two alternatives were compared for each factor.

Two literature reviews on European sediment data and approaches to approximate USLE factors had been conducted. From the alternative USLE maps, the SDR were calculated and simple yet common SDR models were used to explain the spatial variability of SDR and SY. The outcomes and the residuals of the different model realisations were compared to evaluate consequences of the USLE parameterisation and to assess model uncertainties and limitations. The geodata was processed with ArcGIS 9.3.1 (Esri 2009). Statistical analyses were performed with R 2.13.1 (R Development Core Team 2011).

5.3.2 Estimating soil loss

Tab. 21 lists the data and algorithms used to approximate alternative R, K, L and C factors of the USLE. Henceforth, a single (lower case) alphanumeric character refers as code to the calculation of these factors. Soil loss maps are abbreviated by combining the codes of all factors (e.g., gkbc). Due to the topographic similarity at large scales (chapters 2 and 3), only one DEM and S factor were used. The approach by Nearing (1997) is most common for the latter. The P factor was ignored because sufficient data and approaches were unavailable.

For the R and C factors, the literature suggested different approximations for northern and southern Europe. They were separated according to the region Diodato and Bellocchi (2010) developed their R factor model for (Fig. 18). In the overlap between 46° N and 47.5° N, the different approaches were distance-weighted.

The USLE maps (raster-cell size of 0.01 km^2) were compared to a European PESERA map with a resolution of 1 km^2 (Kirkby et al. 2004; Kirkby et al. 2008) to assess how the model choice affects the spatial variability of SDR and the applicability of SDR models. Both models consider sheet and rill erosion but not landslides, gully erosion or riverine sediment transport. However, their concept and outcome differ fundamentally. Unlike the

USLE, the PESERA model is process-oriented and estimates the net erosion of hillslopes. Additionally, the PESERA map only covered the EU-27.

Tab. 21: Datasets and approaches to approximate USLE factors and codes for the alternative approximations. More details and references to datasets in the text

Factor	Code	Datasets	Approaches (references)
R	g	GPCC Reanalysis v4 (1961–2007)	Regional relationships to annual rainfall
	2	CCM2 v2.1 (1975–1999)	
K	k	European Soil Database v2	van der Knijff et al. (1999), Grimm et al. (2003); Poesen et al. (1994)
	m		Neitsch et al. (2002), Grimm et al. (2003); Poesen et al. (1994)
L	b	SRTM DEM	Fuchs et al. (2010)
	1		Asselman et al. (2003)
C	c	CLC and GlobCover	Specific C factors for land cover and land use for northern and southern Europe
	d	ditto, land use statistics	
S	-	SRTM DEM	Nearing (1997)
P	-	-	P=1.0

5.3.2.1 R factor

R factors (in $N \cdot h^{-1} \cdot yr^{-1}$) were derived from long-term average precipitation Pr (in $mm \cdot yr^{-1}$). As this approach is common, alternative R factors were not derived with different regression models but from different precipitation datasets. This takes into account that choosing rainfall gauges and interpolation techniques affects the estimation of R factors. Rainfall data was taken from the CCM2 dataset (Vogt et al. 2007) (code 2) and the GPCC Full Data Reanalysis version 4 with 0.5° resolution (DWD 2008; Schneider et al. 2008) (code g). From the GPCC data, a surface raster was interpolated using the “Natural Neighbor” tool in ArcGIS.

Eq. 19 was used to predict R factors in southern Europe (Diodato and Bellocchi 2010). To obtain R factors in northern Europe (R_{North}), empirical relationships for Germany and Austria (R_{D+AT}) (Strauss and Blum 1994; DIN 2005), Poland (R_{PL}) (Licznar 2006), Belgium (R_{BE}) (Verstraeten et al. 2006), Slovakia (R_{SK}) (Maderková and Antal 2009) and the Alsace (R_{Alsace}) in France (Strauss et al. 1997) were area-weighted and extrapolated (Eq. 20, weighting factor f). Elevation as the explanatory variable in the original approach for R_{PL} was replaced by annual precipitation (Eq. 21). For R_{SK}, the three similar equations in Maderková and Antal (2009) were averaged.

$$R_{South} = 0.199 \cdot Pr - 27.8 \quad \text{Eq. 19}$$

$$R_{North} = f_{D+AT} \cdot R_{D+AT} + f_{PL} \cdot R_{PL} + f_{BE} \cdot R_{BE} + f_{SK} \cdot R_{SK} + f_{Alsace} \cdot R_{Alsace} \quad \text{Eq. 20}$$

$$R_{PL} = 0.1131 \cdot Pr + 6.1878 \quad \text{Eq. 21}$$

5.3.2.2 K factor

K factors (in $Mg \cdot ha^{-1} \cdot h \cdot N^{-1}$) were approximated from soil texture, soil organic content, soil crusting and stoniness of the top soil (Tab. 22). The soil properties were taken from the European Soil Database (ESDB) version 2 (EC and ESBN 2004). For each soil typological unit (STU), only dominant properties were used because information on the proportion of secondary properties is unavailable (C. Bosco, pers. comm.). The soil texture class (attribute TEXTSRFDOM) was transformed to sand, silt and clay content according to van der Knijff et al. (1999). K factors were derived from soil texture following the same authors and Torri et al. (1997) (code k). To calculate alternative USLE maps, the K factor of the modified USLE was used (MUSLE, code m) which considers soil texture and soil organic content (attribute OC_TOP) (Neitsch et al. 2002). For volcanic soils (attribute PARMADO), K was set to 0.8 (van der Knijff et al. 1999).

Tab. 22: Attributes in the European Soil Database, the USLE K factors for texture classes and correction factors for soil crusting and soil organic content

TEXTSRFDOM (soil texture)			CRUST (soil crusting)		OC_TOP (soil organic content)	
Value	K (code k)	K[a] (code m)	Value	Factor[b]	Value	Factor[c]
0	0	0	1	0.75	H (>6%)	0.75
9	0	0	2	0.85	M (2-6%)	0.752
1	0.26	0.14	3	1.00	L (1-2%)	0.83
2	0.43	0.19	4	1.15	V (<1%)	0.975
3	0.35	0.46	5	1.25		
4	0.18	0.40				
5	0.07	0.29				

[a] Without organic content, [b] Grimm et al. (2003), [c] Average factor of interval

Both K factors were corrected for soil crusting (attribute CRUST) (Grimm et al. 2003) and stoniness (attribute WRBFU). For stony soils, i.e. Leptosols and Regosols, K factors were reduced by 70% (Poesen et al. 1994). In a few cases, the attribute WRBFU was undefined. Here, the stoniness was derived from the dominant limitation to agricultural use (attribute AGLIM1). Soils were defined as stony if AGLIM1 was either “gravelly”, “stony”, or “lithic”. Finally, the K factors for the soil mapping units (SMU) were calculated by area-weighting the K factors of the underlying STU.

5.3.2.3 L factor

Slope angles (also used for the S factor, Tab. 11, p. 54) were calculated from the SRTM-DEM (Jarvis et al. 2008) with the "Slope Tool" in ArcGIS ($\beta=\beta_{Nbh}$). An empirical relationship between L and slope angle (Eq. 5, p. 56; code b) was compared to a constant erosive slope length of 100 metres (Eq. 6, p. 56; code 1).

5.3.2.4 C factor

The land cover map for the C factor was created from the Corine Land Cover (CLC) 2006 map, and where not available, from CLC 1990, 2000 (BFS and BAFU 1998; EEA 2010c, b, a), and the coarser GlobCover map (ESA 2008). C factors for the land cover classes and crops were taken from the literature (Jäger 1995; Lóczy et al. 1995; Folly et al. 1996; Auerswald and Kainz 1998; Auerswald and Schwab 1999; Brath et al. 2002; Centeri 2002; Šúri et al. 2002; van Rompaey and Govers 2002; Gabriels et al. 2003; Gómez et al. 2003; Strauss and Wolkerstorfer 2003; Yang et al. 2003; Boellstorff and Benito 2005; Jordan et al. 2005; Morgan 2005; Friedli 2006; Rousseva et al. 2006a; Stumpf and Auerswald 2006; Erdogan et al. 2007; Kliment et al. 2007; Strauss 2007; TLL 2007; Bakker et al. 2008; Lastoria et al. 2008; Märker et al. 2008; Pelacani et al. 2008; Ugur Oczan et al. 2008; de Vente et al. 2009; Kouli et al. 2009; Krása et al. 2010).

Tab. 23: C factors for land cover classes. Upper limit for southern Europe, lower limit for northern Europe

Land cover class	C factor	Land cover class	C factor	Land cover class	C factor
1x[b]	0	213[b]	0.05	11[c]	0.18–0.24
21[ab]	0.24–0.32	221[b]	0.5	14[c]	0.3–0.4
22[b]	0.45	222–223[b]	0.4	20–30[c]	0.23–0.3
23[b]	0.01–0.05	231[b]	0.01–0.05	40[c]	0.005–0.008
24[ab]	0.18–0.24	24x[ab]	0.23–0.3	50[c]	0.001
31[b]	0.005–0.008	31x[b]	0.005–0.008	60[c]	0.01
31–51[b]	0.01–0.05	32x[b]	0.01–0.05	70[c]	0.001
11x[b]	0	331–332[b]	0	90–140[c]	0.01–0.05
211[ab]	0.3–0.4	333–334[b]	0.35	150[c]	0.35
212[ab]	0.18–0.24	335, >400[b]	0	>150[c]	0

[a] If not calculated from land use statistics (Tab. 24), [b] Corine Land Cover, [c] GlobCover

Tab. 24: C factors for crops. Upper limit for southern Europe, lower limit for northern Europe

Land use of arable land	C factor	Land use of arable land	C factor
Abandoned land (subsidized)	0.03–0.1	Oilseeds	0.15–0.2
Cereals	0.12–0.3	Potatoes	0.35–0.5
Cotton	0.5–0.6	Pulses	0.15–0.3
Fallow	0.2–0.7	Root crops	0.35
Forage plants (without maize)	0.05	Sunflowers, soy beans	0.35–0.4
Industrial plants	0.3–0.4	Tobacco	0.5–0.6
Maize	0.45	Vegetables	0.4

Tab. 25: Statistical data on land use of arable land

Coverage	Region	Year	Reference
EU27	NUTS[a] 2	1990–2007	Eurostat (2011b)
	NUTS 3[b]	2000–2007	Eurostat (2011a)
Albania	County	2008	INSTAT (2009)
Croatia	NUTS 3	2003	CBS (2003)
Federation of Bosnia and Herzegovina	Canton	2009	FZS (2010)
Serbian Republic	Total	2008	RZS (2009)
Germany	NUTS 3	2007	Statistical offices of the Federation and the Länder (2011)
Liechtenstein	NUTS 3	2009	AS (2010)
Macedonia	NUTS 3	2007	Macedonian State Statistical Office (2010)
Moldova	Country	2006	Statistica Moldovei (2006)
Montenegro	Municipality	2008	MONSTAT (2010)
Serbia	District	2006	SORS (2007)
Kosovo	Country	2008	SOK (2010)
Switzerland	NUTS 3	2007	BFS (2007)
Turkey	NUTS 3	2006	Turkstat (2011)
Ukraine	Oblast	2008	State Statistics Committee of Ukraine (2009)

[a] Nomenclature des unités territoriales statistiques, [b] Germany NUTS 2

The literature review confirmed higher C factors in southern Europe (as defined in section 5.3.2). The C factors for land cover classes and crops varied enormously in northern and southern Europe and were separately averaged (Tabs. 23–24). Alternative C factors for arable land were either constant (code c) or spatially differentiated (code d). The latter approach of area-weighting crop-specific C factors is common but implicitly assumes that the current land-use pattern reflects the long-term crop rotation. Tab. 25 lists the sources from which land use data was taken. Unfortunately, the heterogeneity of the statistical data (in content and spatial resolution) made the C factors less comparable.

5.3.3 Sediment data and catchment delineation

A preliminary database of sediment data was obtained from published papers and reports. This database partly overlapped with the data collected by Vanmaercke et al. (2011). After locating the monitoring gauges and reservoirs and researching their catchment areas, the catchments were derived from the CCM2 dataset (Vogt et al. 2007) whose accuracy hampered the delineation of small catchments. Different sediment yields for similar monitoring periods were averaged. The bed load of rivers was unavailable.

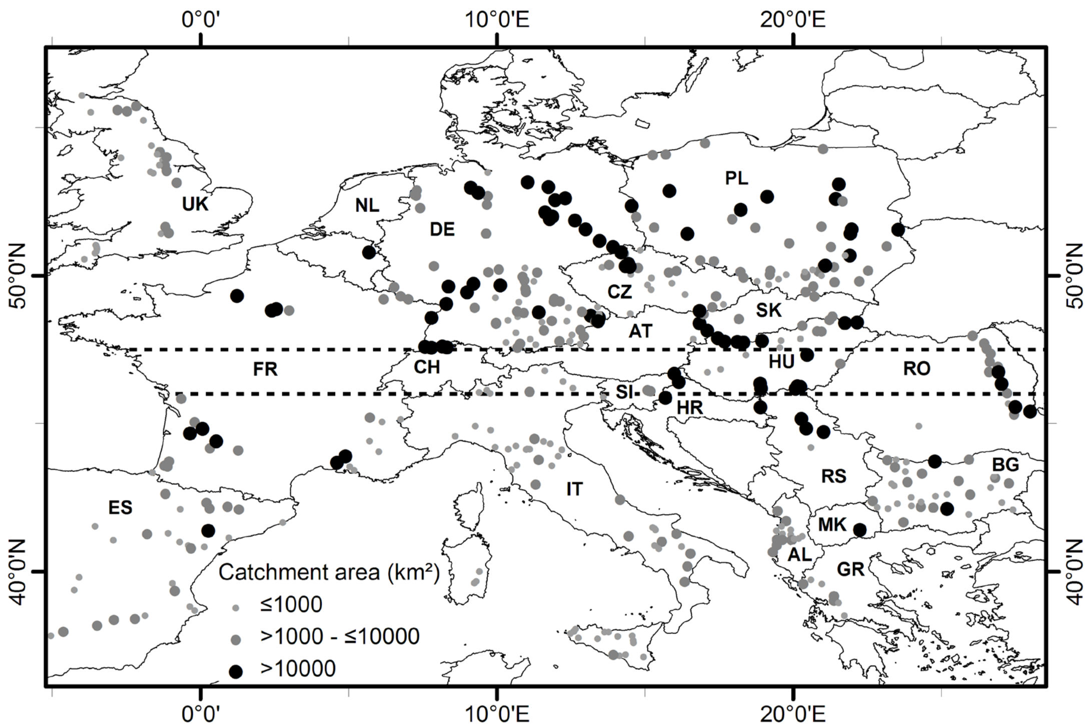


Fig. 18: Selected data of sediment yield. Broken lines indicate the overlap of northern and southern Europe for soil loss modelling. ISO 3166-1 codes for countries with sediment data

The little information on monitoring strategies and load estimation in published data posed a large, unknown uncertainty in sediment yield (cf. Vanmaercke et al. 2011). From the preliminary database, gauges were excluded according to data reliability (e.g., low sampling frequency, undefined location) and deviation of derived to reported (or researched) catchment area (≥25%). The heterogeneous final database comprised more than 400 entries and 10,000 years of riverine and reservoir data across Europe (Tab. 26, Fig. 18).

Tab. 26: Data on sediment yield

Country	No. of data sets (years)	Reference
Albania	16 (362)	Grazhdani (2006); Saraçi (1996)
Austria	1 (10)	Zessner (pers. comm.)
Bulgaria	32 (1084)	Gergov (1996); Rousseva et al. (2006b)
Croatia	1 (163)	Levashova et al. (2004)
Czech Republic	25 (467)	CHMÚ (2006); ICPDR (2004, 2005, 2006, 2007b, a, 2009, 2010b, a); ICPER (2005a, b, 2006, 2007, 2008, 2009, 2010); Kliment et al. (2007); Krása et al. (2005)
France	24 (222)	Asselman (2000); Crouzet et al. (2004); Dumas (2007); Mano et al. (2006); Meybeck et al. (2003); Moatar et al. (2006); Pistocchi (2008); Pont et al. (2002)
Germany	86 (2044)	Asselman (2000); Bechteler (2006); Behrendt et al. (1999); ICPDR (2004, 2005, 2006, 2007b, a, 2009, 2010b, a); ICPER (2005a, b, 2006, 2007, 2008, 2009, 2010); LfU (2009a, b); LUA (2007); LUBW (2009); NLWKN (2011); Pistocchi (2008); Schröder and Theune (1984); Weiss (1996)
Greece	7 (112)	Zarris et al. (2007)
Hungary	21 (225)	Bogárdi (1957); Bogárdi et al. (1983); ICPDR (2004, 2005, 2006, 2007b, a, 2009, 2010b, a); Jordan et al. (2005); Kovács (pers. comm.)
Italy	58 (684[ab])	Diodato and Grauso (2009); Grauso et al. (2008); Torri et al. (2006); van Rompaey et al. (2005); van Rompaey et al. (2003a)
Macedonia	1 (16)	Gobin et al. (2003)
Netherlands	1 (11)	Ward (2008)
Poland	48 (1687)	Brański and Banasik (1996); Łajczak (2003)
Romania	19 (1179)	Levashova et al. (2004); Rădoane and Rădoane (2005); Rădoane et al. (2008)
Serbia	4 (499)	Djorović (1992); Levashova et al. (2004)
Slovakia	8 (44)	SHMÚ (2006, 2007, 2008, 2009, 2011a, b, d, c)
Slovenia	6 (83)	ARSO (2005, 2006, 2007, 2008, 2009); ICPDR (2004, 2005, 2006, 2007b, a, 2009, 2010b, a); Ulaga (2005)
Spain	29 (503[ac])	Batalla and Vericat (2011); Pistocchi (2008); Verstraeten et al. (2003)
Switzerland	1 (14)	Asselman (2000); Pistocchi (2008)
United Kingdom	29 (265[ad])	Brazier (2004); Bronsdon and Naden (2000); Neal et al. (2006); Rowan et al. (1995); Small et al. (2003); Walling et al. (2003); Wass and Leeks (1999)

[a] Long-term reservoir data not included, [b] 24 reservoirs, [c] 17 reservoirs, [d] 12 reservoirs

It consisted of average and annual data which had either been reported as “suspended-sediment yield”, “total suspended solids”, or “sediment yield”. For total suspended solids, the low temporal resolution impeded the estimation of the erosion-independent base load (e.g., industrial effluents, phytoplankton) (section 3.4.2.2). The critical yield as the product of surface runoff and concentration of total suspended solids was available for some gauges in southern Germany (Tab. 10, p. 51). The impact of data disaggregation on lumped SDR models was exemplarily assessed for non-alpine gauges.

5.3.4 Modelling sediment delivery ratios (SDR)

Neither the USLE nor the PESERA model predicts SY. SDR were thus calculated as ratio of sediment yield and soil loss (SDR_{calc}) to adjust the soil loss estimates. For the USLE maps, the SDR subsume both catchment and in-stream retention. As the PESERA model estimates the net soil loss from hillslopes, PESERA-derived SDR are more related to the in-stream retention. To evaluate the impact of algorithm and data choices and assess the applicability of SDR models, four common empirical models of different complexity were applied to regress SDR_{calc} (in %) to catchment properties (Eqs. 22–25).

Of these models, the Area model is the most widely used in erosion studies (cf. de Vente et al. 2007). The Complex model was derived from chapter 3 and recent approaches applied in the Mediterranean zone (Diodato and Grauso 2009) and Europe (Delmas et al. 2009). It reflects the influence of precipitation (Pr), land use and land cover (Ar), topography (β), and drainage density (DD) on sediment connectivity and transport in river catchments. Pr is the long-term average annual precipitation (in $mm{\cdot}yr^{-1}$) and Ar the fraction of arable land (%). The average slope β (in °) is related to other common topographic parameters such as the fraction of flat areas (Delmas et al. 2009). DD ($km{\cdot}km^{-2}$) is the ratio of stream length (in km) to catchment area A (in km^2) and is related to the transport distance of sediments and the fraction of the catchment effectively connected to streams (Delmas et al. 2009). ζ_{11}–ζ_{41} and η_1–η_4 are empirical coefficients.

The CLC classes 21–24 and 211–244 (except 23 and 231) and the GlobCover classes 11–20 were counted as arable land. The input data for Pr in the Complex model was identical to the R factor, i.e. Pr was either derived from the CCM2 or the GPCC dataset (section 5.3.2.1). The stream length was taken from the CCM2 data.

Complex model	$SDR = \eta_1 \cdot \beta^{\zeta_{11}} \cdot DD^{\zeta_{12}} \cdot Pr^{\zeta_{13}} \cdot Ar^{\zeta_{14}}$		Eq. 22
Slope-arable model	$SDR = \eta_2 \cdot \beta^{\zeta_{21}} \cdot (Ar + 20)^{\zeta_{22}}$	Venohr et al. (2011)	Eq. 23
Slope model	$SDR = \eta_3 \cdot \beta^{\zeta_{31}}$	e.g., Diodato and Grauso (2009)	Eq. 24
Area model	$SDR = \eta_4 \cdot A^{\zeta_{41}}$	e.g., de Vente et al. (2007)	Eq. 25

5.3.5 Statistical analyses

The consequences of algorithm and data choice were assessed for the i) soil loss, ii) SDR and iii) SY as product of SDR and soil loss. Ratios and correlation coefficients were calculated between alternative maps to quantify quantitative and qualitative impacts. Kendall's τ was used as correlation coefficient because the data was not normally distributed. High τ values indicate a low impact on SDR and SY models because of spatial similarity.

For SDR and SY, the regression models were evaluated with the Nash-Sutcliffe model efficiency (ME) and adjusted r^2. ME range from $-\infty$ to 1 with higher values corresponding to more accurate and more suitable models. If ME>0, the variance of model residuals is lower than the variance of observed data. The model sensitivity was assessed as change in ME and r^2. The more noticeable the change, the higher the model sensitivity is. Given the different distribution of SDR and SY, deviating SDR did not necessarily made the modelled SY fit the observed values. SY were thus adjusted with a linear regression model ($y=a_1x+a_0$) prior the analyses. The model uncertainty in USLE results and modelled SY was calculated as ratio of the maximum and minimum value.

Tab. 27: Ad-hoc criteria to exclude gauges and catchments from regression analyses

Criteria	Threshold	Explanation
Fraction of arable land	<10%	USLE not explaining important soil erosion processes and sediment sources (channel erosion, glacier erosion, mass movements), low soil loss and high SY led to very high SDR (i.e. leverage values)
Annual sediment yield	>700 Mg·km^{-2}	
Fraction of open area and maximum elevation	>5% and >1,500 m	
SDR_{calc}	≥10,000%	
Data age	≥30 years	Sediment data not reflecting current soil loss
Monitoring period	<5 years	Average SY not representative for long-term conditions
PESERA coverage	<75%	Modelled soil loss not representative

The heterogeneous sediment database consisted of catchments of different sediment yield, environmental settings, and sediment sources as well as monitoring periods and data age. Ad-hoc criteria were defined to reduce inconsistencies between SY and USLE as well as PESERA estimates (Tab. 27). Catchments where processes outside the scope of both models are very likely and unreliable sediment data were excluded. The exclusion criteria and their thresholds were refined evaluating the residuals and outliers during the regression analyses. Common large residuals and influential values mark constraints of the empirical modelling framework and its evaluation. As they can also mask differences between the model realisations, such values were excluded from the statistical analyses while maintaining the sample size as large as possible. As the PESERA map was not everywhere available, catchments with an overlap below 75% were considered as insufficient for the statistical analyses. However, they were used to compare USLE alternatives.

Tab. 28: Ad-hoc definition of (sub-)regions. Catchments associated to the country their gauge falls in

Region	Sub-regions	Countries
North (n=149)	DE (n=52)	Germany
	PL (n=47)	Poland
	W (n=19)	France, western Germany, Netherlands, United Kingdom
	-	Austria, Croatia, Czech Republic, Hungary, Macedonia, Serbia, Slovakia, Switzerland
Central-South East (CSE, n=30)	-	Austria, Croatia, Czech Republic, Hungary, Macedonia, Romania, Serbia, Slovakia, Slovenia, Switzerland
South (n=95)	SW (n=25)	France, Spain
	SE (n=47)	Albania, Bulgaria, Greece, Hungary, Romania
	IT (n=23)	Italy
	-	Croatia, Macedonia, Serbia, Slovenia

Taking the remaining unexplained variability into account, the model quality and its sensitivity to the soil loss estimation was also evaluated in several regions (Tab. 28). How do the SDR models perform in different parts of Europe? For which soil loss map are the results acceptable (ME>0)? Are the applicability and constraints of the empirical modelling framework and the sensitivity of ME and r^2 to the soil loss and SDR models consistent? To answer these questions, the gauges were grouped by country and associated to the regions North, South, and, as transition region, Central-South East (CSE). In general, the northern part of Europe (i.e. north of the Alps) is characterised by low to moderate SY in flat and high SY in alpine areas. The climate is humid with equally-distributed annual rainfalls. In southern Europe (mainly the Mediterranean zone), SY are higher and the climate features a pronounced seasonality with dry summers and wet winters. Gully erosion, not covered in the USLE and PESERA, can be an important sediment source (Vanmaercke et al. 2012a). The large regions North and South were exemplary sub-divided.

5.4 Results and discussion

5.4.1 Soil loss maps

The soil loss estimates and model uncertainty varied widely between European regions (Figs. 19–20). Both are not related. For the three regions in Fig. 20, the methods for estimating C and K factors influenced the USLE results the most. Their average uncertainty was 60%, unlike the low uncertainty of the R factor (<10%). The average maximum model uncertainty (of all USLE factors) was >100% with values ranging from 0% to >1,000%. Unlike the USLE, the PESERA results were similar or, in southern Europe, even below SY. These findings are in accordance to Vanmaercke et al. (2012a).

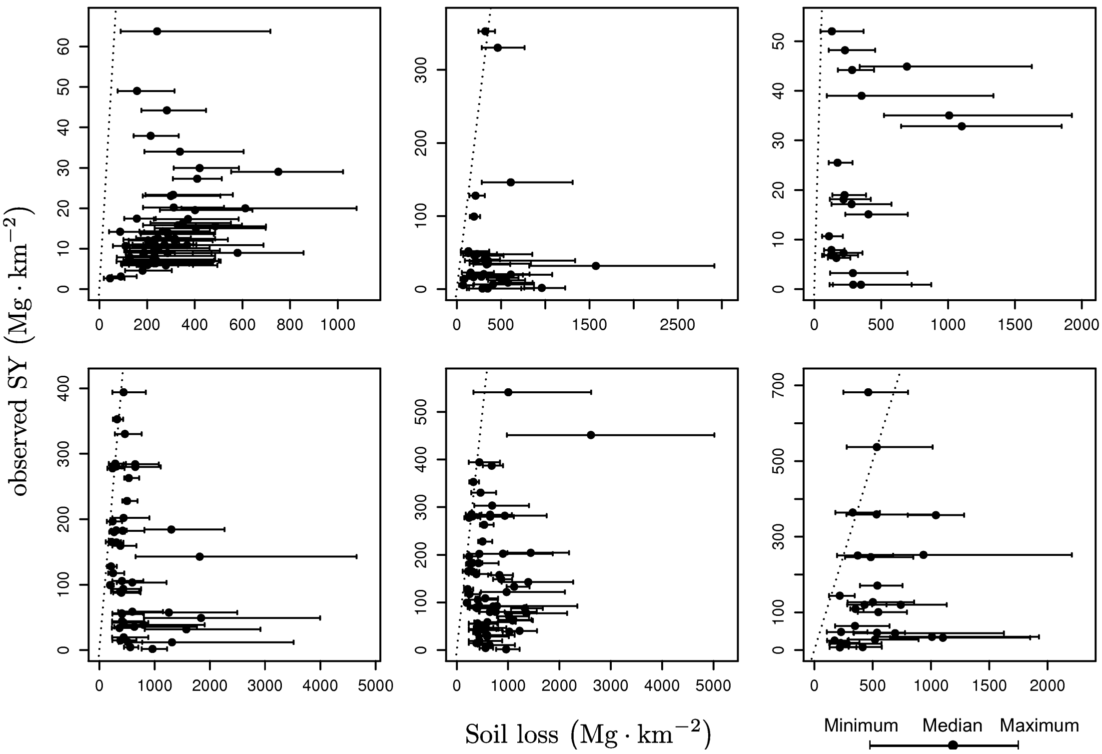


Fig. 19: Sediment yield and uncertainty in USLE estimates. Top: sub-regions DE, PL, and W of region North, bottom: region CSE, sub-regions SE and SW of region South

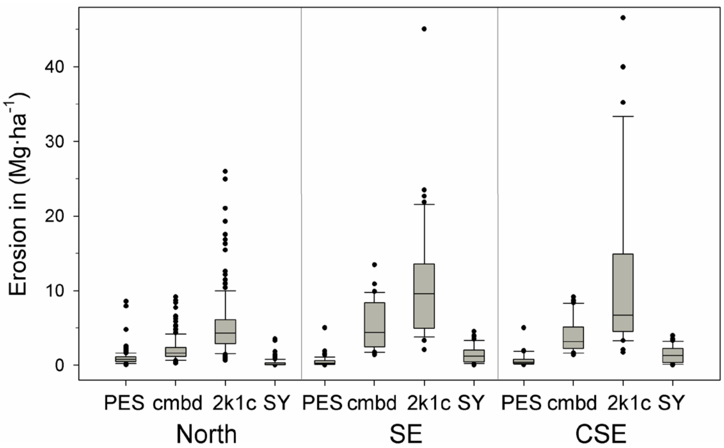


Fig. 20: Soil loss and sediment yield for catchments in European regions (Tab. 27). PES is the PESERA map, 4-letter codes for USLE maps (Tab. 21), SY is the sediment yield

Consequently, choosing the soil loss model and approximation approaches for model parameters caused not only an enormous uncertainty in absolute values of SDR_{calc} but, more important for analysing the sensitivity of SDR models, also affected the spatial pattern of SDR_{calc}. The correlation between all USLE maps and the PESERA map was loose. Kendall's τ ranged from -0.3 to 0.1 for the regions in Fig. 20 and from 0.0 to 0.25 for the whole dataset. The correlation coefficients were significantly higher among the USLE maps with τ being maximal between opposite R and C factors (τ>0.8). For the K and L factors, the algorithm choice affected the spatial pattern of the soil loss, and thus SDR, more strongly. For contrasting USLE maps, the average correlation coefficient τ was about 0.6 (0.45–0.72) for the complete dataset and varied from 0.36 to 0.82 among the regions.

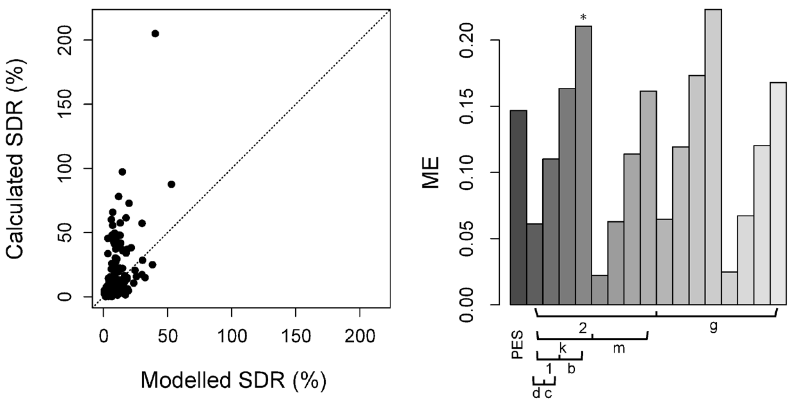


Fig. 21: Quality and sensitivity of the Complex model in region North. Model predictions with the USLE map 2kbc (left), model sensitivity to soil loss estimation (right, 2kbc marked with an asterisk), PES is the PESERA map, alphanumeric characters refer to the alternative USLE factors (Tab. 21)

5.4.2 Sediment delivery ratio (SDR)

The SDR models were insufficient to explain the spatial variability of SDR of the pre-selected river catchments (Fig. 21). Despite the ad-hoc criteria, influential values strongly affected the regression analyses. High SDR_{calc} and residuals were found to be associated to the criteria defined in Tab. 27. They point to sediment sources not considered in the soil loss models and are discussed as model and data limitations in the (sub-) regions.

5.4.2.1 Region North

5.4.2.1.1 Sub-region DE

The Complex model (Eq. 22) well explained the SDR variability with some USLE maps (Tab. 29). While the Slope-arable model (Eq. 23) was also, albeit to a lesser degree, acceptable, the Slope and Area models (Eqs. 24–25) could not adequately describe the spatial pattern at all. The Complex and Slope-arable models were sensitive to the approximation of the K factor and (for USLE maps) the SDR parameterisation. The C and L factors were less important and the R factor negligible. For the Complex model, constant C factors for arable land (code c) were on average slightly more favourable than variable C factors (code d). Accordingly, USLE maps like gmbc allowed the best model agreement, while the opposite map 2k1d gave poor results.

Changing the soil loss model also affected significantly the model performance. Compared to the best USLE maps, the predictive power of the Complex and Slope-arable models was low if applied to SDR calculated with the PESERA map. The impact was stronger than switching from the Complex to the Slope-arable model.

Tab. 29: Model performance for sub-region DE. Results for alternative R factors were similar (codes g and 2). Best model performance in bold type

Soil loss map	n	Complex model ME	Complex model r^2	Slope-arable model ME	Slope-arable model r^2	Slope model ME	Slope model r^2	Area model ME	Area model r^2
gk1d	44	0.18	0.09	0.05	0.02	-0.0	0	0.05	0.06
gk1c	44	0.45	0.35	0.06	0.02	-0.0	0	0.07	0.13
gkbd	44	0.46	0.42	0.18	0.25	0.15	0.15	-0.0	0
gkbc	44	**0.59**	**0.52**	0.13	0.16	0.06	0.08	-0.0	0.05
gm1d	44	0.53	0.42	0.38	0.36	-0.0	0	-0.0	0
gm1c	44	**0.65**	**0.50**	0.34	0.30	-0.0	0	-0.0	0
gmbd	44	**0.64**	**0.60**	**0.42**	**0.53**	0.18	0.16	-0.0	0
gmbc	44	**0.69**	**0.60**	0.37	0.42	0.09	0.09	-0.0	0
PESERA	43[a]	0.20	0.38	0.17	0.30	0.18	0.27	0.06	0.14

[a] Leverage point not considered

The analyses of the whole dataset (n=52) was strongly affected by leverage points (e.g., SDR_{calc}>250% for the PESERA map). All SDR models underestimated the same high SDR_{calc} with every soil loss map due to important sediment sources outside the scope of the modelling framework. These values were excluded from the regression analyses. This was not only relevant for the influential Alpine river Iller (SY=63 $Mg{\cdot}km^{-2}{\cdot}a^{-1}$). Considerable underestimations also occurred for other high SY (≈35 $Mg{\cdot}km^{-2}{\cdot}yr^{-1}$) measured at the only reservoir and at river Itz. This data was considered unreliable because of old age (reservoir, 30 years old) and short monitoring period (1991–2002, river Itz). Additionally, the model deviations were also large for the lowland river Ems with average SY (≈10 $Mg{\cdot}km^{-2}{\cdot}a^{-1}$) where human activity like shipping traffic and intensive agriculture, leading to eutrophication and phytoplankton growth, made the measured SY overestimate the contribution of sheet and rill erosion (Fuchs et al. 2012).

5.4.2.1.2 Sub-region PL

In contrast to sub-region DE, the slope-based SDR models equally well explained the spatial variability of SDR (Tab. 30). The applicability of the SDR models was most sensitive to the L factor and the soil loss model. Satisfactory results were obtained with the empirical L factor (code b) but not with constant erosive-slope lengths (code 1). Model results were much better with the PESERA map than with any USLE map. Other USLE factors and especially the parameterisation of the SDR model were less important.

Tab. 30: Model performance for sub-region PL. Results for alternative R factors were similar (codes g and 2). Best model performance in bold type

Soil loss	n	Complex model		Slope-arable model		Slope model		Area model	
map		ME	r^2	ME	r^2	ME	r^2	ME	r^2
gk1d	29	0.16	0.09	0.09	0.13	0.10	0.15	-0.0	0
gk1c	29	0.26	0.15	0.14	0.17	0.14	0.19	-0.0	0
gkbd	29	**0.49**	**0.44**	**0.46**	**0.46**	**0.48**	**0.47**	-0.0	0
gkbc	29	**0.56**	**0.47**	**0.49**	**0.48**	**0.50**	**0.50**	-0.0	0
gm1d	29	0.06	0.17	0.05	0.17	0.04	0.11	-0.0	0
gm1c	29	0.12	0.18	0.10	0.18	0.10	0.14	-0.0	0
gmbd	29	0.36	0.38	0.40	0.36	0.39	0.30	-0.0	0
gmbc	29	0.42	0.38	0.45	0.37	0.43	0.32	-0.0	0
PESERA	27[a]	**0.80**	**0.81**	**0.79**	**0.82**	**0.79**	**0.83**	-0.1	0

[a] Two catchments outside PESERA map

The residuals were large for catchments with high SY and SDR due to model constraints. Similar to DE, the highest values (SY>70 $Mg \cdot km^{-2} \cdot a^{-1}$) occurred in the mountainous southern Poland where landslides and channel incision are important sediment sources (Lach and Wyżga 2002; Rejman and Rodzik 2006). Additionally, the abundant heavy industry and mining activity in the Upper Silesian Coal Basin caused high loads of anthropogenic suspended matter in the lowland river Przemsza (Helios Rybicka 1996; Dulias 2010). Consequently, and similar to the German river Ems, there was a strong mismatch in the flat terrain (β<2°) between low modelled SDR (ζ_{11}, ζ_{21}, and ζ_{31}>0) and low soil loss rates, i.e. high SDR_{calc}.

5.4.2.1.3 Sub-region W

The Complex and Slope-arable models gave acceptable results with some USLE maps, while the two simpler models failed (Tab. 31). The predictability of SDR was sensitive to the choice of the R and L factors. For the Complex model, the soil loss model was similarly relevant. In contrast to the other sub-regions, the GPCC data (code g) and constant erosive-slope lengths (code 1) were much more favourable than their alternatives. Apart from the C factor when using the L factor code b, other choices were unimportant for the model quality.

Catchments at the rivers Lowman and the Cadière were removed prior the regression analyses because they had sparse arable land (Ar<17%) yet maximal SY (SY≈50 Mg·km^{-2}·a^{-1}). Important natural and anthropogenic sediment sources were not covered here by the modelling framework (cf. Fiandino and Martin 2004; Walling and Collins 2005). The comparatively low modelled soil loss led to influential SDR_{calc} and partially to very high ME and r^2. The model performance of the Complex model was better (0.17≤ME≤0.93) compared to simpler models (-0.14≤ME≤0.64 for the Slope-arable model).

Tab. 31: Model performance for sub-region W. Results for R factor code g were better than for code 2. Best model performance in bold type

Soil loss map	n	Complex model		Slope-arable model		Slope model		Area model	
		ME	r^2	ME	r^2	ME	r^2	ME	r^2
gk1d	17	**0.25**	0.58	**0.16**	0.63	-0.2	0	-0.2	0.18
gk1c	17	**0.22**	**0.67**	**0.14**	**0.69**	-0.1	0	-0.3	0.27
gkbd	17	0.13	0.56	-0.0	0.59	-0.2	0	-0.2	0.28
gkbc	17	-0.2	**0.64**	-0.3	**0.63**	-0.3	0	-0.3	0.34
gm1d	17	**0.20**	0.57	0.07	0.59	-0.1	0	-0.3	0.05
gm1c	17	**0.23**	**0.68**	**0.12**	**0.65**	-0.1	0	-0.4	0.13
gmbd	17	0.16	0.58	-0.0	0.58	-0.1	0	-0.3	0.11
gmbc	17	-0.0	**0.67**	-0.2	**0.62**	-0.2	0	-0.4	0.17
PESERA	17	0.09	0.21	**0.12**	0.32	-0.0	0.11	-0.1	0

5.4.2.1.4 Entire region

After removing the catchments which showed large residuals in the sub-regions and harmonising the exclusion criteria (reservoir data, SY>40 Mg·km^{-2}·a^{-1} and Ar<20%), the final dataset (n=107) covered 12 out of the 14 countries of region North. The model performance was only acceptable with the Complex and Slope-arable models (0.04≤ME≤0.26). The sensitivity to single USLE factors was weak. The best results were obtained with USLE maps based on the combination of constant C factors for arable land, empirical L factors and the MUSLE K factor (codes c, b, and m). The impact of the soil loss model, SDR parameterisation and R factor was negligible. Similar to sub-regions DE and PL, the USLE maps 2mbc and gmbc allowed the best model predictions.

5.4.2.2 Region CSE and sub-region SE

Without the influential but unreliable reservoir data of old age (≥24 a) and short monitoring period (5 a), no model could describe the SDR variability in region CSE. Strong differences occurred when only catchments covered by the PESERA map were considered (n=19). For these, positive ME in region CSE were obtained with the Complex but not with the simpler models (Tab. 32). The model outcome was thus most sensitive to the SDR parameterisation. Applying the same exclusion criteria, no model was adequate to explain the SDR variability in sub-region SE (ME<0).

Tab. 32: Model performance for region CSE. Results for R factor code 2 were slightly better than for code g. Best model performance in bold type

Soil loss map	n	Complex model ME (ME^b)	Complex model r^2	Slope-arable model ME (ME^b)	Slope-arable model r^2	Slope model ME (ME^b)	Slope model r^2	Area model ME (ME^b)	Area model r^2
gk1d	25	-0.1 (0.35)	0.32	-0.2 (-0.4)	0.02	-0.4 (-0.4)	0	-0.4 (-0.4)	0.04
gk1c	25	-0.1 (0.42)	0.36	-0.2 (-0.4)	0.01	-0.4 (-0.4)	0	-0.4 (-0.4)	0.04
gkbd	25	-0.2 (0.33)	0.35	-0.2 (-0.3)	0.06	-0.4 (-0.3)	0	-0.4 (-0.4)	0.09
gkbc	25	-0.1 (0.40)	0.39	-0.2 (-0.3)	0.05	-0.4 (-0.3)	0	-0.4 (-0.4)	0.09
gm1d	25	-0.2 (0.37)	0.30	-0.2 (-0.3)	0.03	-0.4 (-0.4)	0	-0.4 (-0.4)	0.04
gm1c	25	-0.1 (0.45)	0.34	-0.2 (-0.3)	0.02	-0.4 (-0.4)	0	-0.4 (-0.4)	0.05
gmbd	25	-0.1 (**0.47**)	0.34	-0.2 (-0.3)	0.09	-0.4 (-0.3)	0.03	-0.4 (-0.4)	0.08
gmbc	25	-0.0 (**0.54**)	0.37	-0.1 (-0.2)	0.08	-0.4 (-0.3)	0.02	-0.4 (-0.4)	0.09
PESERA	19[a]	(0.42)	0.57	(-0.2)	0.14	(-0.2)	0.19	(-0.4)	0.02

[a] 6 catchments outside PESERA map, [b] Harmonised dataset (n=19)

All models systematically underestimated the SDR of Romanian catchments in both regions while overestimating the SDR elsewhere. These catchments belong to the Siret river basin in eastern Romania. Geology, climate and land use in the mountainous setting favour moderate to severe soil erosion by water including gully erosion and sediment yields which are among the highest in Romania (Ionita et al. 2006; Rădoane et al. 2008; Zaharia et al. 2011). Nonetheless, the modelled soil loss was comparatively low resulting in above-average SDR (Tab. 33). For the Romanian catchments, the Complex model predicted the SDR well with USLE maps based on the MUSLE K factor (code m) and the PESERA map (0.64≤ME≤0.71, 0.50≤r^2≤0.65) but less so with the alternative K factor (0.35≤ME≤0.47, 0.04≤r^2≤0.38). Accordingly, the sensitivity to this factor was strongest. The sensitivity to the SDR parameterisation was comparatively low. The Slope-arable model behaved almost equally in terms of model performance and sensitivity. Acceptable results were also possible with the simple SDR models in combination with codes m and 1 (Area model) and the PESERA map (Slope model).

Tab. 33: Mean soil erosion rates (range) of catchments in regions CSE and SE (without excluded values)

Value	Romania (n=12)	CSE – Romania (n=12)	SE – Romania (n=27)
Minimum soil loss (USLE) ($Mg{\cdot}km^{-2}{\cdot}a^{-1}$)	244 (116–451)	516 (227–821)	597 (234–987)
Maximum soil loss (USLE) ($Mg{\cdot}km^{-2}{\cdot}a^{-1}$)	571 (390–1075)	2192 (633–4656)	1473 (512–5019)
SY ($Mg{\cdot}km^{-2}{\cdot}a^{-1}$)	228 (118–394)	51 (4.5–184)	97 (4.5–451)
Minimum SDR (%)	42 (26–64)	2.5 (0.3–8.1)	6.4 (0.6–14)
Maximum SDR (%)	108 (55–191)	9.5 (1.0–23)	17 (1.0–46)

For the non-Romanian catchments in region CSE, however, the model sensitivity to the USLE parameterisation was weaker in comparison to the SDR parameterisation and the soil loss model. The gross overestimation of the unusually low SDR_{calc} of the small Kali basin at Lake Balaton (A=82 km^2) affected the model performance of the Complex model. The range of SDR_{calc} of this catchment (0.1–0.2%) was also much lower than the values reported by Jordan et al. (2005) based on a distributed erosion and sediment transport model as well as detailed input data (1.2–1.4%) implying far too high soil loss estimates. In contrast, other low SDR_{calc} of nearby catchments at rivers Wulka in Austria ($1.5<SDR_{calc}<3.5\%$) and Zala in Hungary ($0.6<SDR_{calc}<1.1\%$) were found to be consistent with literature values (Kovacs et al. 2012). Without the Kali basin, all USLE maps allowed almost equally well predicted SDR with the Complex model ($0.5 \leq ME \leq 0.6$, $0.41 \leq r^2 \leq 0.56$). The model quality was poorer with simpler models ($ME \leq 0.17$). For the PESERA map, switching from the Complex to the Slope-arable model had no impact on the model applicability (ME=0.27). However, the simpler models were not suitable ($ME \leq 0$).

Different results were observed for the non-Romanian catchments in sub-region SE, after removing four extreme values including the Kali basin. The high SY of the river Topolnitsa (300 $Mg{\cdot}km^{-2}{\cdot}a^{-1}$) is strongly affected by mining (Tz. Karagiozova, pers. comm.) (cf. Bird et al. 2009). A similar discrepancy of sparse arable land (Ar<20%), i.e. low modelled soil loss, and huge SY (390 $Mg{\cdot}km^{-2}{\cdot}a^{-1}$) occurred at the river Luda Kamchiya in Bulgaria. Although it can be assumed that sediment sources other than sheet and rill erosion are important in many Bulgarian catchments (Rousseva et al. 2006b), the modelled soil loss and observed SY disagreed by far the most in these two mountainous catchments. One exceptionally low value from river Maritsa was considered as not representative. The short monitoring period (1980–89) was characterised by severe droughts and drastically diminished sediment transport throughout Bulgaria (Tz. Karagiozova, pers. comm.).

The approximation of the K and L factors had the largest impact among the USLE factors. ME were above 0.6 (Complex and Slope-arable models) and ranged from 0.4–0.5 (Slope model) when the MUSLE K and the empirical L factor were used or between -0.1 and 0.46 otherwise. Switching to the PESERA map decreased the model performance even more (ME=0.17 for the Complex, ME≤0.07 for the Slope-arable and Slope models). The SDR parameterisation was far less relevant and negligible for the most suitable maps 2mbc, 2mbd, gmbc, and gmbd. The Area model was not appropriate at all.

5.4.2.3 Region South

The SDR models could not explain the large variability of SDR in the region South. The model performance was poor (ME≤0). The SDR models could only be applied successfully in sub-regions.

5.4.2.3.1 Sub-region IT

The maps gmbc and 2mbc were most appropriate to predict the SDR of Italian catchments (Tab. 34). The performance of the slope-based SDR models was acceptable while the Area model failed. The model quality was sensitive to the soil loss model and USLE factors except the R factor.

Tab. 34: Model performance for sub-region IT. Results for alternative R factors were similar (codes g and 2). Best model performance in bold type

Soil loss map	n	Complex model ME (ME^b)	r²	Slope-arable model ME (ME^b)	r²	Slope model ME (ME^b)	r²	Area model ME (ME^b)	r²
gk1d	22	-0.0 (-0.0)	0	-0.0 (-0.0)	0	-0.1 (-0.1)	0.0	-0.1 (-0.1)	0
gk1c	22	-0.0 (0.02)	0	0.0 (0.04)	0.03	-0.0 (-0.1)	0.0	-0.1 (-0.1)	0
gkbd	22	0.43 (0.21)	0.27	0.41 (0.15)	0.28	0.42 (0.07)	0.27	-0.1 (-0.0)	0
gkbc	22	0.45 (0.23)	0.30	0.49 (0.21)	0.32	0.46 (0.09)	0.28	-0.1 (0.02)	0
gm1d	22	0.41 (0.40)	0.20	0.27 (0.32)	0.19	0.23 (0.02)	0.13	-0.0 (0.21)	0.1
gm1c	22	0.54 (0.48)	0.27	0.36 (0.42)	0.25	0.25 (0.05)	0.14	0.08 (0.28)	0.11
gmbd	22	0.63 (0.61)	0.44	0.51 (0.49)	0.45	0.51 (0.29)	0.40	-0.1 (0.20)	0.03
gmbc	22	**0.77 (0.67)**	**0.50**	**0.65 (0.60)**	**0.50**	**0.62 (0.33)**	**0.42**	-0.1 (0.26)	0.05
PESERA	18[a]	(0.29)	0.27	(-0.2)	0	(-0.2)	0	(-0.2)	0

[a] Leverage points not considered (SDR_{calc}>1,300%), [b] Harmonized dataset (n=18)

The models could not explain the average SDR_{calc} of the catchment of Lago di Pontecosi with the steepest average gradient (β=18°) and the least fraction of arable land (Ar=13%). The mismatch is related to constraints of the soil loss models, although sheet and rill erosion contribute in general only partly to the high SY of the Italian catchments (49–650 Mg·km^{-2}·a^{-1}, median 237) (cf. de Vente et al. 2006).

5.4.2.3.2 Sub-region SW

In the sub-region SW, the outcome of the Complex model was acceptable while the simpler SDR models were not adequate (with USLE maps; Tab. 35). Apart from the strong sensitivity to the SDR parameterisation, the results were sensitive to the soil loss model and the K but not the other USLE factors. The best results were achieved with USLE maps based on the K factor code k while its alternative was less suitable. For the Complex model, the impact of the soil loss model was similar to the K factor. Better results were obtained with USLE maps. In contrast, the PESERA map was much more suitable for predicting SDR with the Slope-arable model.

Tab. 35: Model performance for sub-region SW. Left: whole sub-region, right: Spanish data. Results for R factor code 2 were slightly better than for code g. Best model performance in bold type

Soil loss	n	Complex model		Slope-arable model		n	Complex model		Slope-arable model	
map		ME	r^2	ME	r^2		ME	r^2	ME	r^2
2k1d	22	0.42	0.55	-0.0	0.11	16	**0.26**	**0.23**	0.11	0.15
2k1c	22	**0.46**	**0.65**	-0.1	0.16	16	**0.28**	**0.25**	0.11	0.14
2kbd	22	0.42	0.64	0.02	0.27	16	0.17	0.17	-0.0	0.06
2kbc	22	**0.47**	**0.73**	-0.0	0.30	16	0.20	0.19	-0.0	0.05
2m1d	22	0.28	0.43	-0.0	0.11	16	0.22	0.19	0.14	0.23
2m1c	22	0.30	0.52	-0.1	0.14	16	0.23	0.20	0.14	0.22
2mbd	22	0.26	0.49	-0.0	0.21	16	0.14	0.06	0.05	0.10
2mbc	22	0.28	0.58	-0.1	0.24	16	0.15	0.07	0.05	0.09
PESERA	22	0.33	0.62	**0.29**	0.34	16	0.29	0.14	0.21	0.08

Besides river Cadière in southern France (section 5.4.2.1.3), the suspiciously low SY measured at the Spanish reservoir La Tranquera were not used in the analyses. All regression models overestimated its exceptionally low SDR_{calc} (1.4–2.7%). However, the observed 8 $Mg{\cdot}km^{-2}{\cdot}a^{-1}$ had been attributed to either measurement errors or human activity in the catchment (Sanz Montero 2002). Additionally, the simple modelling framework did not consider important sediment sources (gully and bank erosion) and controlling factors (grazing intensity, protective land cover) of the SY pattern (de Vente et al. 2005). The mismatch of high SY and low modelled soil loss led to the very high SDR_{calc} of some Spanish catchments. The underestimation of these SDR was strongest for the reservoir Fuensanta (SY=680 $Mg{\cdot}km^{-2}{\cdot}a^{-1}$), whose mountainous catchment had the least arable land in sub-region SW (Ar=12%). To emphasize the differences between the alternative models, this common and influential outlier was not considered.

The SY of Spanish catchments (27–540 $Mg·km^{-2}·a^{-1}$) were measured in reservoirs and were much higher than the French river data (SY<45 $Mg·km^{-2}·a^{-1}$). Despite the homogeneous measurement technique in Spain, the predictive power of the SDR models was lower than in the region SW given the model limitations. The sensitivity of the regression models to the SDR and USLE parameterisation was similar yet less pronounced.

5.4.3 Sediment yield (SY)

For analysing the sensitivity on SY, problematic and influential catchments were not considered. The SDR models have been applied to the refined regions of the previous section. Due to the poor prediction of SDR in most regions, the Slope and Area models are not addressed in this section.

5.4.3.1 Region North

The Complex and Slope-arable models predicted well the variability of SY in the different parts of region North (Tab. 36). The adjustment of the modelled SY was only significant when applying the PESERA map and the R factor code g in the sub-region W. However, the model quality and sensitivity of SY estimates to the soil loss estimation differed not only between the regions but also between SDR and SY. In general, ME and r^2 were higher and the model sensitivity lower for SY than for SDR. Furthermore, the relative impact of soil loss, SDR parameterisation and USLE approximation differed.

The SDR models equally well explained the spatial pattern of SY in the sub-region PL. Unlike SDR, the SY prediction was not sensitive to the L factor of the USLE. Accordingly, the remarkably good SDR results with the PESERA map did not positively affect the prediction of SY. In sub-region DE, however, the model sensitivity was more pronounced, especially for the Slope-arable model. Similar to the SDR prediction, the MUSLE K factor (code m) was more suitable than its alternative. The impact was stronger for the Slope-arable than for the Complex model. Likewise, the sensitivity of the SDR model to the L and C factors was negligible for the Complex model but not the Slope-arable model. In contrast to the SDR, the model agreement was better with variable C for arable land (code d) and constant erosive-slope lengths (code 1) instead of their alternatives. However, both SDR models were most sensitive to the soil loss estimation. Using the PESERA map instead of the USLE maps led to much lower ME, which could not be compensated by adjusting the modelled SY.

The soil loss model was also most important for the model applicability in sub-region W. Again, the results were better with the USLE maps than with the PESERA map. ME and r^2 were higher with USLE maps based on the R factor code 2 – in contrast to the SDR results. The discrepancy was especially large for the unadjusted

Complex model (-0.2<ME≤0.5 with code 2 and ME<0.1 with code g). Unlike the SDR, the model quality was also sensitive to the USLE K factor. The influential deviation of one catchment was reduced with the adjustment.

Tab. 36: Modelled SY in region North. Results for R factor code 2 similar to code g for sub-regions DE and PL and better for sub-region W. Best model performance in bold type

Soil loss map	DE				PL				W			
	Complex		Slope-arable		Complex		Slope-arable		Complex		Slope-arable	
	ME	r^2	ME	r^2	ME	r^2	ME	r^2	ME[ab]	r^2	ME	r^2
gk1d	0.57	0.56	0.53	0.52	0.77	0.76	0.73	0.72	**0.61**	**0.58**	**0.62**	**0.60**
gk1c	0.60	0.59	0.37	0.35	0.81	0.81	0.76	0.76	**0.59**	**0.57**	**0.64**	**0.61**
gkbd	0.61	0.60	0.42	0.41	**0.84**	**0.83**	**0.82**	**0.82**	0.55	0.52	0.58	0.55
gkbc	0.57	0.56	0.23	0.21	**0.87**	**0.86**	**0.84**	**0.83**	0.54	0.51	0.60	0.57
gm1d	**0.75**	**0.74**	**0.71**	**0.70**	0.70	0.69	0.75	0.74	0.51	0.47	0.51	0.48
gm1c	**0.72**	**0.71**	0.53	0.52	0.74	0.73	0.77	0.77	0.47	0.43	0.47	0.43
gmbd	**0.72**	**0.72**	**0.62**	**0.62**	0.76	0.75	**0.82**	**0.81**	0.45	0.42	0.46	0.43
gmbc	0.65	0.64	0.42	0.41	0.78	0.78	**0.82**	**0.82**	0.42	0.38	0.41	0.37
PESERA	0.19	0.17	0.13	0.11	0.77	0.77	0.69	0.76	0.32	0.28	0.28	0.24

[a] Offset of linear SY adjustment of SY differs significantly from 0 (p<0.05), [b] Only significant for R factor code g

Both SDR models also predicted sufficiently the variability of SY in the entire region North with some USLE maps (0.2<ME≤0.42, 0.2<r^2≤0.41 after adjustment). Two catchments commonly deviated: the Kali basin whose low SY was grossly overestimated (see section 5.4.2.2) and a catchment in southern France. Even though the residuals increased for SY above 30 $Mg{\cdot}km^{-2}{\cdot}yr^{-1}$, the large deviation of the French catchment (SY=32 $Mg{\cdot}km^{-2}$) was attributed to the short monitoring period of 5 years. Without both catchments, the model performance improved (0.30≤ME≤0.48 after adjustment).

In accordance to the sub-regions DE and W, the PESERA map was less favourable than the USLE maps. In fact, the model performance was most sensitive to the soil loss model followed by the R and C factors. The C factor code c and R factor code g allowed better model results than their alternatives. Considering the prediction of SDR and SY, no soil loss map can be considered as “most suitable” everywhere. Nonetheless, the maps 2kbc in combination with the Complex model allowed acceptable results in all parts of region North. Compared to the other maps, the explained variability in the entire region North was also among the highest (r^2=0.37, ME=0.38).

5.4.3.2 Regions CSE and South

Applying the SDR models to predict the SY variability of catchments in southern Europe was of limited use (ME<0). Satisfactory results were only possible for sub-regions (Tab. 37). In comparison to region North, the sensitivity to the soil loss model was similar but to the SDR model more pronounced. The model quality and explained variability dropped considerably when the PESERA instead of USLE maps was used. Moreover, ME were negative with the PESERA map.

For the Romanian catchments in regions CSE and SE, both SDR models allowed satisfactory SY predictions (ME=0.55-0.72 for the Complex, ME=0.45-0.64 for the Slope-arable model). The model sensitivity to USLE factors was accordingly low. The best results were achieved with the maps 2m1c and gm1c. The SDR models were again most sensitive to the soil loss model. The results were poor with the PESERA map (ME=r^2=0).

Tab. 37: Modelled SY in regions CSE, SE and SW. Results for R factor code 2 differed slightly from code g. Best model performance in bold type

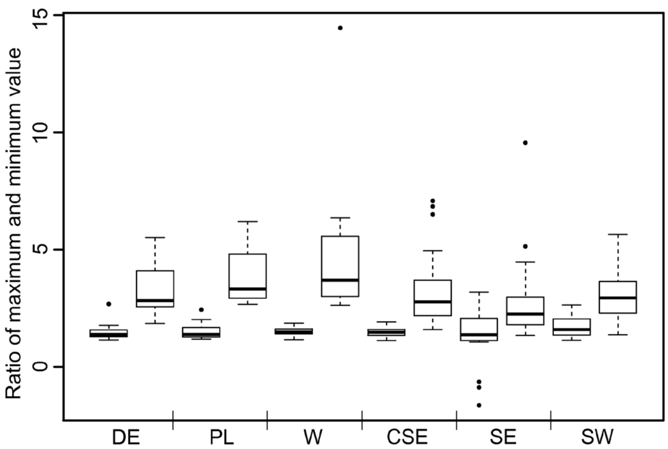
Soil loss map	CSE (without Romania)				SE[a] (without Romania)				SW			
	Complex		Slope-arable		Complex		Slope-arable		Complex		Slope-arable	
	ME	r^2	ME	r^2	ME	r^2	ME	r^2	ME	r^2	ME	r^2
gk1d	**0.77**	**0.74**	0.30[b]	0.23	0.39	0.36	0.30	0.27	0.27[cd]	0.24	0.18[c]	0.14
gk1c	**0.79**	**0.77**	0.30[b]	0.23	0.42	0.40	0.32	0.29	0.28[cd]	0.24	0.14[bc]	0.10
gkbd	0.66	0.62	0.20[b]	0.12	0.53	0.52	0.45	0.42	**0.43**	**0.40**	**0.26**	**0.23**
gkbc	0.73	0.70	0.22[b]	0.14	0.56	0.55	0.46	0.44	**0.43**	**0.40**	0.18[bdc]	0.14
gm1d	**0.81**	**0.79**	**0.43**	**0.37**	0.52	0.50	0.44	0.42	0.23[cd]	0.19	0.16[bc]	0.12
gm1c	**0.79**	**0.77**	**0.39**	**0.33**	0.55	0.54	0.46	0.44	0.23[cd]	0.19	0.13[bc]	0.08
gmbd	0.75	0.73	0.36	0.29	**0.62**	**0.60**	**0.54**	**0.52**	0.25	0.22	0.17[bdc]	0.12
gmbc	**0.79**	**0.77**	0.36	0.30	**0.64**	**0.63**	**0.56**	**0.54**	0.25[cd]	0.21	0.12[bc]	0.08
PESERA	0.04[b]	0	0.05[b]	0	0.16[bc]	0.12	0.21[c]	0.17	**0.41**[c]	**0.38**	**0.28**[c]	**0.24**

[a] Without one leverage value (outside the PESERA map), [b] Slope of linear SY adjustment differs not significantly from 0 (p<0.05), [c] Offset of linear SY adjustment differs significantly from 0 (p<0.05), [d] Only significant for R factor code g

The results for the non-Romanian catchments in CSE and SE slightly differed. The combination of the Complex model and USLE predicted the variability of SY in region CSE much better than alternative combinations. Although the results were better with the MUSLE K factor (code m) instead of code k (before the adjustment), the sensitivity was low to the USLE parameterisation compared to the SDR model. In contrast, satisfactory results for region SE were obtained equally well with both SDR models (ME≤0.78). However, the evaluation for region

SE was affected by an influential value (SY=450 $Mg{\cdot}km^{-2}{\cdot}a^{-1}$). Without it, the impact of changing the SDR model as well as the K and L factors became more distinct while ME and r^2 decreased (Tab. 37).

The model performance for catchments in the region SW corresponded to the SDR results. Without the linear adjustment, the Complex model partly allowed acceptable results (ME≤0.39) while the Slope-arable model almost failed to capture the regional SY variability (ME≤0.12). To obtain acceptable model results, the USLE parameterisation was more important than the choice of the SDR model. In contrast to the SDR prediction and other (sub-)regions, all ME were negative except for USLE maps based on the empirical L factor code b and the K factor code k. For Spanish catchments, the model quality was even lower (ME<0.3). Large residuals of the highest SY and SDR (section 5.4.2.3.2) were decisive in Spain and the region SW. The negative ME revealed that the calibrated SDR models could not sufficiently explain the SY with the PESERA map without linear adjustment. Only after the adjustment, the model performance was comparable to the USLE maps. In Italy, however, the results only partly reflected the model performance for SDR predictions. They were best for the combination of MUSLE K (code m) and constant C factors for arable land (code c; ME≤0.28 without adjustment). The alternative K factor code k was less appropriate. The sensitivity to the L factor and the soil loss model was low compared to the SDR prediction. Apart for the most suitable soil loss maps (gm1c, 2m1c, PESERA), the SDR model hardly affected the model performance.



Value		North			
		All	DE	PL	W
Max / Min	SY	1.3	1.4	1.4	1.5
	Soil loss	3.3	2.8	3.3	3.7
Coefficient of variance	SY	0.1	0.1	0.1	0.1
	Soil loss	0.4	0.3	0.4	0.4
		South and CSE			
		CSE	SE	SW	
Max / Min	SY	1.5	1.4	1.6	
	Soil loss	2.8	2.3	2.9	
Coefficient of variance	SY	0.1	0.1	0.2	
	Soil loss	0.3	0.3	0.4	

Fig. 22: Uncertainty in modelled sediment yield and soil loss. Left: SY (left boxes) and soil loss (right boxes), right: median values, SY calculated with Complex SDR model

The results confirmed the findings for northern Europe that the empirical SDR models were in general less adequate to relate PESERA estimates to the observed SY. They also showed that the Complex model was more appropriate for southern Europe than the Slope-arable model. Again, no USLE map was "best" for all regions.

Integrating the model performance for SDR and SY, model results were on average most apprpriate with the maps gmbc, gmbd, 2mbc, and 2mbd in the sub-regions of CSE and SE.

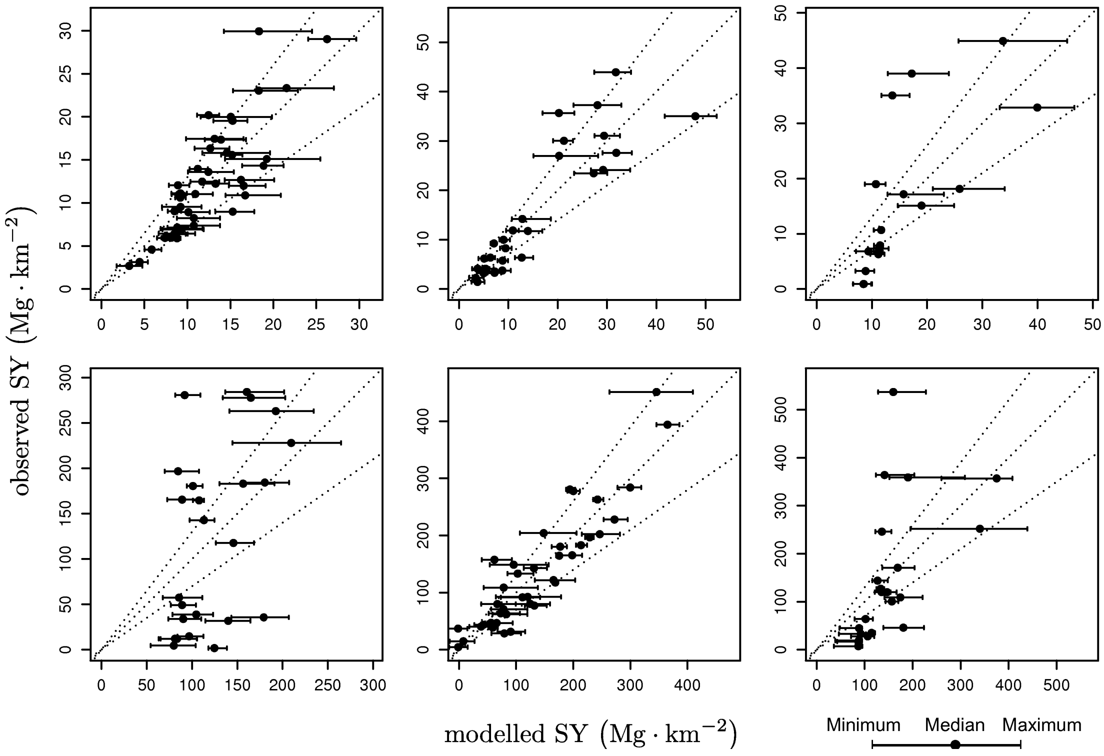


Fig. 23: Uncertainty in modelled sediment yields and observed values (top: DE, PL, W, bottom: CSE, SE, SW). Modelled with the Complex model, dotted lines are the 1:1 line ±30%.

5.4.3.3 Uncertainty in soil erosion

As expected, the uncertainty in modelled SY was significantly lower than the uncertainty in soil loss ($p<0.001$; Fig. 22). It ranged on average from about 30% to 60% in the different regions but was considerable larger for individual catchments, especially in southern and western Europe (Fig. 23). The average model uncertainty had thus the same order of magnitude as the uncertainty in the literature values (Fig. 24 left) and in sediment data (e.g., Walling and Webb 1981; Vanmaercke et al. 2011).

The potentially large impacts of low sampling interval and data interpolation have been discussed in sections 1.2.3 and 3.5.1.2. Another source of uncertainty is the unknown erosion-related fraction of total SY in rivers. The available critical fractions were on average about 1/3 of total SY (Fig. 24 right). The significant correlation ($p<0.001$) suggests a minor relevance for the spatial pattern of SDR and SY. Empirical SDR models could simply be adjusted. However, the results indicate more variability for SY>30 $\mathrm{Mg \cdot km^{-2}}$. For the two highest values (>40 $\mathrm{Mg \cdot km^{-2}}$), though, other discharge-dependent processes are presumably dominant in the catchments (in-stream erosion, urban runoff) which could not be separated from sheet and rill erosion (section 3.5.2) thus supporting the refined exclusion criteria for region DE and North.

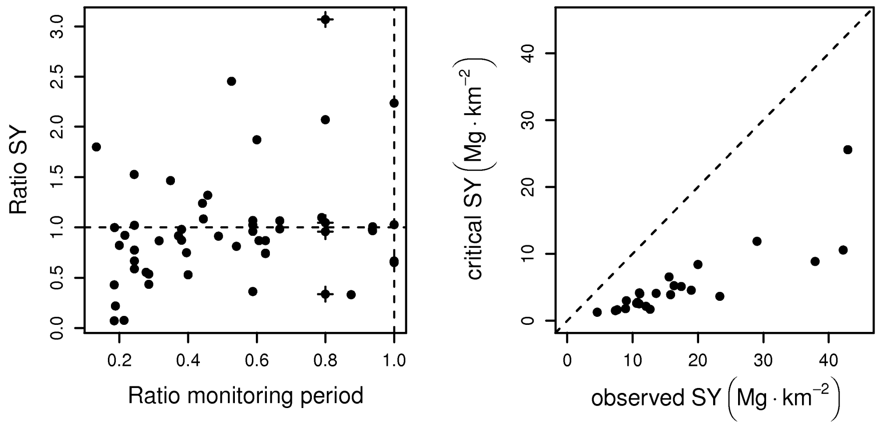


Fig. 24: Uncertainty in sediment data. Left: Variability of reported sediment yields, ratios of values from different data sources shown if the monitoring period fell inside a longer monitoring period, monitoring periods estimated for reservoirs in Spain (reported as “several decades”) (stars), right: Estimated sediment yield of surface runoff (critical SY) and observed values for non-alpine catchments in southern Germany

5.5 Discussion – Application and evaluation of SDR models

The findings show that simple empirical SDR models can sufficiently explain the spatial variability of SDR and SY of European river catchments. As expected, the soil loss and SDR models have a strong impact on the model applicability. Among the tested alternatives, combining the USLE and the Complex model based on topography, drainage density, land cover, and precipitation allowed the most satisfactory results. Empirical SDR models with fewer parameters are less applicable in southern Europe. Diodato and Grauso (2009), with more detailed soil loss data and a few catchments, as well as Pistocchi (2008), with different sediment data and SDR models, made similar observations. Of the three alternative SDR models, the Slope-arable model is almost equally applicable in northern Europe while the Area and Slope models cannot be recommended. The poor results with the PESERA map reveal that the fundamentally different concept behind PESERA requires other catchment features to link its outcomes to observed sediment yields.

Besides highly uncertain soil loss and significantly less yet considerably uncertain modelled SY, the uncertainty in approximating USLE factors for European soil loss maps also affects the performance of SDR models. In northern Europe, the impact is stronger than the impact of changing from the Complex to the Slope-arable model. None of the approaches can be recommended to optimally predict SDR and SY everywhere. However, the map 2kbc is proposed as most suitable for predicting SY in European regions, although the K factor code k was less favourable for predicting SDR and the sensitivity to the R factor was usually low.

The favourable approximation of USLE factors at large scales is “more suitable” rather than “more precise” and, without doubt, the uncertainty goes far beyond two alternative approximations. The results were repeatedly similar or even better with constant C factors for arable land, although variable values might be considered as better conforming to the variability of soil loss in river catchments. In either case, average values (for C factors and crops) only crudely reflect the heterogeneity of the C factor. The same holds true for the other USLE factors. Therefore, the study can shed only some light into the uncertainty and sensitivity of SDR models at large scales. Nonetheless, the findings clearly show that carefully choosing the approaches to estimate USLE factors can positively affect the performance of SDR models. Therefore, conceptually new approaches to obtain USLE factors like interpolating local K factors instead of using the coarse soil classes of the European Soil Database (Panagos et al. 2012) offer not only the possibility to describe the variability of soil erodibility differently (if not better) but also to further improve the applicability of empirical SDR models.

The results also show that model quality and sensitivity differ between SDR and SY. For instance, the sensitivity to the soil loss model was much stronger when predicting SY instead of SDR. SY predictions can also be acceptable with poorly predicted SDR, contrary to what might be expected. In addition, model uncertainty and sensitivity are variable in space. SDR models should thus be evaluated against SY and SDR in different regions. Furthermore, the variable impacts of algorithm and data choices on model quality and sensitivity support the hypothesis that regional findings cannot be simply extrapolated.

Despite the uncertainty in model parameterisation, the replacement of USLE maps to obtain more satisfactorily applicable SDR models is limited. In the present study, none of the evaluated algorithm and data choices overcomes some general mismatches between the modelling framework and the sediment yield as integral of all sediment sources. For instance, high soil erosion rates occurring in Mediterranean catchments prone to gully erosion (García-Ruiz 2010; Cantón et al. 2011) are only marginally related to USLE and PESERA estimates. In agreement to previous studies (van Rompaey et al. 2003b), only very narrow solutions for SDR models can be expected. Other catchment features like the abundance of gullies and landslides have been shown to be better predictors for the high SY in countries like Italy and Spain (de Vente et al. 2005; de Vente et al. 2006). Other plausible reasons for influential model deviations in all parts of Europe require more specific knowledge about river catchments. This covers natural processes in areas with sparse arable land ($Ar \leq 20\%$) like bank and moorland erosion and anthropogenic factors like mining, water diversion and sediment trapping in reservoirs.

The huge uncertainty in the heterogeneous sediment data (Vanmaercke et al. 2011) has an unknown effect on the applicability, uncertainty and sensitivity of SDR models. Important sources are related to the temporal variability of soil loss and sediment transport. For instance, SY are more uncertain and less representative for long-term average conditions if measured over short monitoring periods (Vanmaercke et al. 2012b). The SY of different age also vary considerably (Fig. 24 left) and are less comparable. Information on the monitoring period (and thus data age) is often readily available and should be used to qualitatively assess the reliability and comparability of sediment data. However, other vital information about sampling frequency and data interpolation are rarely provided.

The topic of data comparability also comprises how SY are measured (e.g., in suspension, as sedimentation rate). In the present study, data was measured as either "suspended solids" or "suspended sediments" with very few exceptions. Although using total suspended solids arguably overestimates the contribution of soil erosion to SY, the significant correlation between erosion-related and total suspended solids in southern Germany suggests that SDR models could simply be adjusted to erosion-related yields in regional studies but more detailed data is needed for verification. In contrast, the SDR of reservoirs did not fit the SDR of river catchments. The few reservoir data was usually very high and old compared to riverine data. However, the poor model performance in both SW Europe (riverine and reservoir data) and Spain (only reservoir data) has been primarily the consequence of limited soil loss estimates rather than incommensurable sediment data. Similarly, the repeatedly large residuals of small catchments of Hungarian reservoirs might equally be explained by the data age, the measurement technique, (ephemeral) gullies as important sediment sources (Podmanicky et al. 2011), or by the combination of the three.

In fact, in many catchments several model and data constraints are likely to superimpose making it difficult to pinpoint the reason for influential values and residuals. For instance in the UK (sub-region W), the USLE maps were derived with empirical R factors from Central Europe leading to a potential overestimation of R factors due to the high annual rainfall. Most data has also been sampled during the national Land-Ocean Interaction Study (LOIS) programme during the 1990s. Therefore, the sediment yield is probably not representative for long-term average conditions. Additionally, the sediment yield of catchments in flat terrain with sparse arable land ranges from below to above average (1–60 $Mg{\cdot}km^{-2}{\cdot}a^{-1}$). While low SY are in agreement with low USLE S and C factors, the high SY suggest other important sources of (suspended) sediments (e.g., Walling et al. 2002).

The broad range of natural and artificial conditions and the unknown uncertainty in the heterogeneous sediment data hamper the calibration and evaluation of any large-scale erosion model. Sediment budgets of catchments and techniques like source fingerprinting and un-mixing models are already used to break up the black box “sediment yield” (Walling et al. 2011; Hinderer 2012). A sufficient number of such sediment budgets across Europe would certainly be helpful to overcome some problems of calibrating and assessing empirical SDR models in the future.

5.6 Conclusions

The many approximations of USLE factors used in the past in large-scale studies, the different (potential) soil loss maps make the uncertainty in pan-European soil loss estimates large and, moreover, highly variable. Additionally, new input data and erosion-related maps are constantly prepared. The resultant uncertain spatial pattern of calculated sediment delivery ratios (SDR) raises the question how the application and evaluation of empirical erosion models are affected. The empirical model character and the limited input data do not allow judging any alternative model realisation to be more precise in describing the soil erosion in river catchments. The intention of this study can only be to quantify the model uncertainty and sensitivity to model, algorithm and data choices and assess the suitability of potential soil loss maps for a specific task, namely for predicting SDR and sediment yields of European river catchments.

Unsurprisingly, the model concept for soil loss is found to generally exert a strong impact on the performance of SDR models. In contrast to USLE maps, the different SDR models do not fit the concept of the process-oriented PESERA model. Other explaining factors are required to link sediment yields to PESERA estimates. In accordance to previous studies, SDR models with more parameters are found to be more favourable to predict the SDR and SY of European catchments. The model performance was (much) better with drainage density, topography, precipitation and land use as model parameters than with fewer parameters such as catchment area, topography, or a combination of topography and land use. Nonetheless, the quality of a given SDR model and the explained variability of SDR and sediment yield are also sensitive to the approximation of USLE factors. The sensitivity varies among different regions and is in the same order of magnitude as the SDR parameterisation. Among the tested alternatives, the K factor (soil erodibility) is generally most important, followed by the C and L factors (land cover, erosive-slope length). The tested different input data for the R factor (rainfall) is mostly less relevant. However, no soil loss map can be unanimously recommended as “the best” map for predicting SDR and sediment yields with empirical SDR models.

Besides the sensitivity of the model performance, algorithm choices are also a significant source of uncertainty in modelled sediment yield. Calibrated SDR models thus require a minute documentation of the underlying soil loss map to avoid misapplications. The model uncertainty is similar to the variability of sediment yields in the literature and is arguably comparable to the uncertainty in observed sediment data and the unknown base load of total suspended solids. For the latter, further studies are needed to verify the significant correlation between total and critical yields observed in southern Germany.

However, all model realisations fail for the huge variability of SDR and sediment yield across Europe. Data and model constraints only allow satisfactory regional models. Large residuals usually occur if sediment data is unreliable and not comparable to other data and if the soil loss models do not capture important erosion processes and sediment sources. Such data and catchments have to be removed to avoid negative impacts of outliers and influential residuals on the sensitivity analyses. However, the range of erosion processes, human activity, and environmental conditions make it difficult to link modelled soil loss to literature values of sediment yield. The simple model concept, error-prone input data and the heterogeneous sediment data blur the distinction between model errors and inappropriate sediment data.

Although additional information on sediment processes outside the scope of the USLE and PESERA models, new approaches to better reproduce the variability of USLE factors, more parameters in the SDR model to cover (artificial) sediment trapping, and homogeneous sediment data to reduce inconsistencies are needful to overcome limitations in model applications, algorithm choices are expected to remain influential for the performance of SDR models. Carefully choosing the soil loss map contributes to the successful application of SDR models at large scales. For the sensitivity analyses, however, well-distributed and representative data is strongly recommended as local findings are not valid for whole countries or river basins.

6 Summary

Modelling the variability of soil erosion in space and time is an important part of assessing mass transfers in river catchments and coupling the fluxes of nutrients and contaminants with in-stream processes. The complexity of the phenomenon "soil erosion" promoted the development of many models in the past. Choosing a model depends on its purpose and data availability among other criteria. At large scales like nations and river basins, empirical models are used to predict the spatial and temporal variability of soil loss, sediment delivery ratios (SDR) and sediment yield from limited input data. It is good practice in any modelling – apart from striving for better predictions – to evaluate the uncertainty of model outcomes, sensitivities to input data and algorithms, and to identify model limitations. This dissertation thesis addresses all four topics for empirical models in four studies in the European context because two extensive literature reviews confirmed the use of various data and algorithms in Europe to derive erosion-related model parameters and the high variations of sediment yield. For the first time, algorithm and data choices in modelling soil erosion at regional and European scales are broadly and systematically analysed. The major findings provide new insights into the relevance of choices for (critical) sediment yield, topography, and soil loss – fundamental for any modelling and evaluation of soil erosion –, into the application and evaluation of empirical soil erosion models as well as their improved yet limited applicability at large scales.

First, data and algorithm choices for topographic parameters and critical yields of suspended solids are found to exert significant uncertainties in the parameterisation of erosion models and in calibration and validation data (chapters 2 and 3). The raster resolution of digital elevation models (DEM) is most influential. Especially for coarse DEM, algorithm choices are also important even for basic parameters like the slope angle. The combined effect of uncertainty in approximated factors of the universal soil loss equation (USLE) makes the uncertainty in European soil loss maps and, subsequently, in modelled sediment yields even larger (chapter 5). The findings show model uncertainties to be in the same order of magnitude as the uncertainty in measured sediment yields. A minute documentation of calibrated models is thus required to avoid misapplications.

Second, the sensitivity of models to topographic uncertainty is nonetheless low (chapters 2 and 3). Therefore, more detailed DEM cannot improve explaining the regional variability of soil erosion in river catchments. These results extend and generalize findings of Pan and King (2011) as well as Yong et al. (2009). In contrast, artefacts can hamper the higher data resolution (chapter 2; de Vente et al. (2009)). As long as the DEM resolution is insufficient for GIS-based algorithms, local regression models to predict slope-length (L) factors of the USLE are recommended (chapter 3). Unlike topography, the estimation of USLE factors and soil loss significantly influences the model performance (chapter 5). Most important for USLE maps is the soil (K) and, to a lesser degree, the land management (C) and L factors. However, no "best" map for all regions in Europe could be identified. For the estimation of critical yields, the findings are mixed. In contrast to the spatial variability of topographic parameters, the prediction of the annual variability of critical yields of suspended solids benefits from higher data resolution and the careful method choice respectively (chapter 3). The functional streamflow disaggregation approach (FSD) with its daily time step makes the inter-annual variability of critical yields more plausible than the statistical approach based on (multi-)annual average values. However, the spatial model performance is worse. The significant differences for average critical yields at some monitoring gauges require further analyses.

Third, although careful algorithm choices can thus improve the performance of SDR models, they are certainly not sufficient to overcome the observed model limitations. The extensive sediment database established for this dissertation thesis clearly shows the constraints and limitations of the regional and European application of empirical SDR models. These limitations are essentially related to the discrepancy between the processes and sources contributing to the observed and the modelled soil erosion (chapters 3 and 5). Specifically, the USLE restricts the prediction of sediment yields with SDR models to non-alpine catchments outside the Mediterranean zone. The enormous temporal variability of soil erosion not only impedes the prediction of the extraordinarily high sediment yields during flood years in alpine catchments (chapters 3 and 4) but also the use of unrepresentative average sediment yields (i.e. old data, short monitoring period) for model calibration and validation (chapters 3 and 5). Without complementing the USLE and better disaggregating the sources of sediments and suspended solids, regional sediment data has to be used to calibrate and/or evaluate regional SDR models. However, modelling all (relevant) soil erosion sources and processes at large scales and separating the sediment from the modelled sources and processes is far from being solved.

Fourth, the combination of pan-European data, USLE and SDR models can satisfactorily predict the variability of sediment yields in space and time – despite the uncertainty and within the limitations mentioned above (chapters 3–5). Nonetheless, two model improvements are proposed. Chapters 3 and 4 show that surface runoff estimated with the FSD approach and seasonally weighted rainfall data are more suitable than simple annual rainfall indices to explain the inter-annual variability of sediment yields. As daily data of water discharge needed for FSD is not available everywhere, the generic, parsimonious RAMSES model evaluated in chapter 4 is suggested for large-scale applications. Additionally, chapters 3 and 5 highlight that calibrating SDR models to sediment yields does not necessarily lead to good SDR predictions and poor SDR predictions do not, in turn, mean poor predictions of sediment yield. SDR models should therefore always be calibrated and evaluated against both SDR and sediment yields. Doing so for catchments in southern Germany, it is found that existing SDR models without hydrological parameters cannot explain the spatial variability of SDR, despite satisfactory results for sediment yields (chapter 3). Considering hydrology as a new parameter in SDR models is helpful to significantly improve the model quality. In chapter 5, the 4-parameter SDR model derived from chapter 3 (Eq. 22) is successfully applied in different parts of Europe. Its predictions of SDR and sediment yields are generally superior to SDR models with fewer parameters (including the one used in the MONERIS model), especially outside Central Europe. However, even with various soil loss maps, no empirical SDR model is capable of explaining the enormous variability of SDR and sediment yields in Europe. This result is in agreement to former studies (e.g., Verstraeten et al. 2003; de Vente et al. 2007). Reliable sediment data, regionally optimised USLE factors, and recognizing non-USLE sediment sources are expected to improve future applications.

The extensive analysis in chapter 5 proposes a new USLE map for predicting sediment yields of European river catchments. Based on this map (code 2kbc in chapter 5), a new 4-parameter SDR model is derived, valid for catchments north of the Alps where USLE processes are in general the major contributors to sediment yields. Although it is less suitable for explaining the variability of SDR, Eq. 26 satisfactorily predicts the sediment yield of catchments with more than about 20% arable land. For sub-regions, however, specific regression models are recommended. With the on-going research to improve large-scale soil loss maps (Cerdan et al. 2010; Meusburger et al. 2012; Panagos et al. 2012), Eq. 26 should be regularly re-evaluated. For the Mediterranean zone, however, the combination of USLE and SDR models is of limited use due to the poor model performance for catchments with very high SDR values. One reason is that the USLE does not consider gully erosion which is an important sediment source in less vegetated areas (e.g., Martín-Fernández and Martínez-Núñez 2011).

$$\begin{aligned} SDR &= 0.4776 \cdot \beta^{-0.498} \cdot DD^{0.953} \cdot Pr^{1.244} \cdot Ar^{-1.245} \\ SY &= 0.780 \cdot SDR \cdot E + 4.285 \end{aligned} \qquad \text{Eq. 26}$$

Eq. 26 predicts the long-term SDR and sediment yield (SY) from common catchment properties such as topography (β), drainage density (DD), average annual rainfall (Pr), and land use (Ar). However, two related problems have to be kept in mind when using Eq. 26 (and any other SDR model). First, the SDR values for establishing Eq. 26 are derived from total sediment yields of river catchment. This empirical model thus does not describe the sediment flux into rivers. Second, the soil loss, sediment delivery (ratio), and in-stream sediment retention cannot be independently validated with one sediment yield measured at the outlet. Without additional data, any model evaluation is restricted to how well the variability of SDR and sediment yield is explained and how suitable soil loss maps for this task are. Promising steps to overcome model limitations and to (further) improve model predictions are discussed in the last chapter.

7 Outlook

Some research topics can be immediately derived from the previous chapter in order to improve empirical estimations of the variability of soil erosion in river catchments and broaden the evaluation of model uncertainties and sensitivities. They comprise both the estimation of the critical fraction of total suspended solids as well as model parameters (Tab. 38). Given the huge variability of environmental settings, soil losses, sediment yields, and SDR in Europe, an inter-regional focus is recommended.

Tab. 38: Overview of proposed further research topics

Research area	Topics
Temporal variability (RAMSES model, Eq. 13)	Generalisation of the seasonal weighting and application in non-alpine catchments
	Explanatory variables for sediment yields during flood years (in different climate zones)
Disaggregation of suspended-solid data	Critical yields of suspended solids in different parts of Europe (re-calibrating SDR models)
	Assessing the separation of overland flow
	Functional disaggregation of suspended-solids data
Spatial variability	Regional regression models for the USLE R factor (e.g. in NW Europe)
	Evaluating the relevance of measured K factors (Panagos et al. 2012) for SDR models
	Matching soil loss estimations, erosion processes, and sediment yields
Disaggregation of SDR	Separating in-stream and catchment retention

First, neither simple regression models nor the RAMSES approach explained the extraordinary high sediment yields occurring during a few flood years. As most of the sediment transport is mobilised during hours to a few days, aggregated rainfall totals are likely to mask this variability. Pan-European data of higher temporal resolution like the E-OBS dataset on daily rainfall (Haylock et al. 2008) promises to be a starting point not only for a better estimation of USLE R factors (cf. Angulo-Martínez and Beguería 2009) but also for deriving model parameters which are closer related to the annual variability of soil erosion. At first glance, however, annual indices based on the daily rainfall data seem not to be *per se* much more suitable than annual rainfall (Fig. 25). For instance, the number of rainfall events (daily values exceeding a certain threshold) and annual percentiles are well-correlated to annual sums. Although totals of the highest rainfalls and especially the maximum daily rainfall deviate significantly, their variability is inconsistent for years with high soil erosion (Fig. 25c–d).

Furthermore, the huge variability of catchment responses to daily rainfall rates has also to be recognized. For instance, if topsoils are saturated or sediment freshly deposited, subsequent rainfall is likely to trigger higher soil erosion rates than under contrary initial conditions. Fig. 25 also shows the seasonal response which is related to vegetation cover and, in mountainous catchments, to snowfall and melting. Therefore, further studies should focus on combining seasonally weighted daily data (e.g., using the leaf area index) and aggregated data reflecting the overall wetness to explain the inter-annual variability of soil erosion more effectively. Nonetheless, the enormous spatial variability of storm intensities even over short distances (Fiener and Auerswald 2009) – much shorter than the distance between rainfall gauges – will make any large-scale modelling of surface runoff and sediment mobilisation in heterogeneous catchments uncertain.

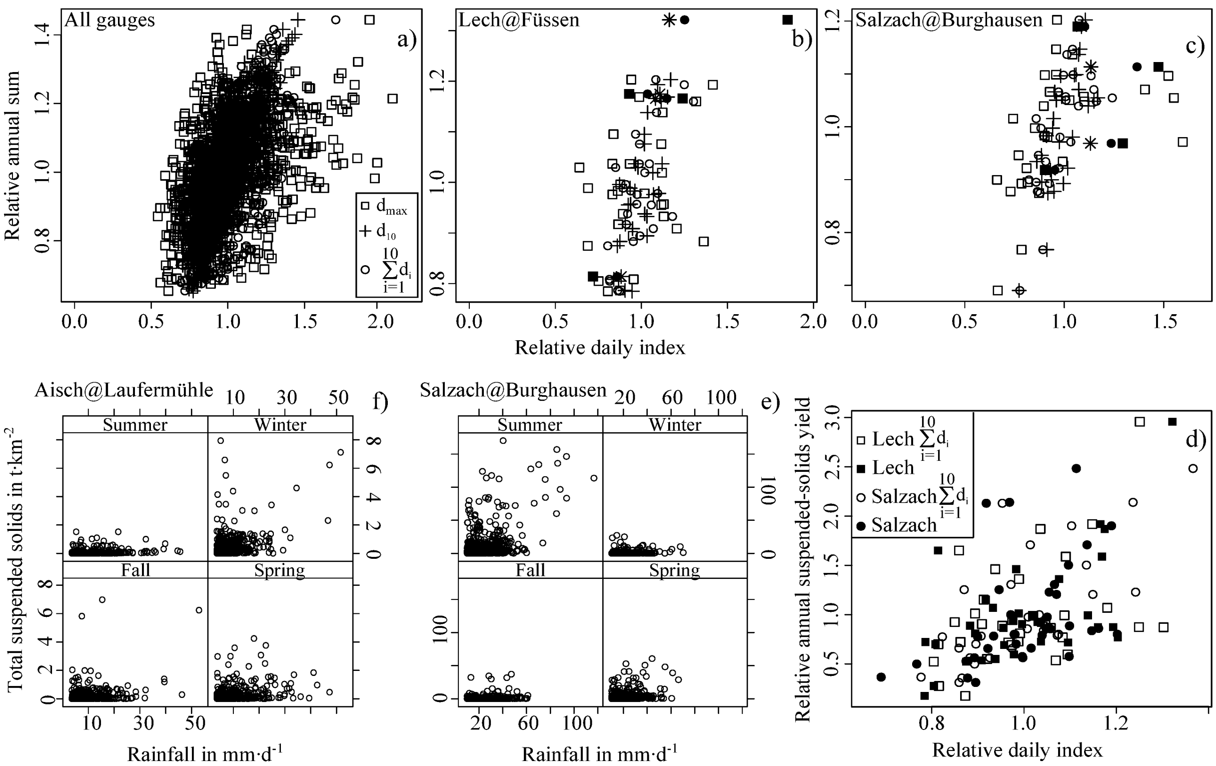


Fig. 25: Relationships of daily rainfall indices to annual rainfall and to suspended-solids yields for exemplary catchments in southern Germany, a) relationships between simple annual indices based on daily rainfall for all catchments (cf. Tab. 10 and section 4.3.1), annual maximum daily rainfall (d_{max}), the annual 97.26th percentile, i.e. the 10th highest value (d_{10}), and the sum above d_{10} (Σd_i), b–c) similar to a) but for two alpine catchments, high annual suspended-solids yields are highlighted, d) gauge-specific relationships between annual rainfall (black), the sum above d_{10} (white), and annual suspended-solids yields, e–f) seasonality of relationships between daily rainfall for rainfall events (>75th percentile) and suspended-solids yields mobilised within 2 days after the rainfall events for an alpine (Salzach) and non-alpine catchment (Aisch). The daily maximum rainfall for the catchments extracted from the E-OBS data v5.0 (Haylock et al. 2008), the daily suspended solids data for 1971–2003 similar to chapters 3 and 4

Second, the promising first application of the "functional streamflow disaggregation" (FSD) to estimate critical suspended-solids yields has to be further scrutinized. The large uncertainty and large residuals for some monitoring gauges in chapter 3 call for case studies with more detailed data to test how reliably FSD separates the "fast" mode. Additionally, the assumption in chapter 3 that this "fast" mode is always equivalent to overland flow needs to be further assessed. For instance, the results for the gauge at Taferlruck in chapter 3 suggest that in mountainous and forested catchments not every fluctuation in water discharge is related to overland flow and soil erosion.

Despite the methodical development and the new software version (Carl 2011), FSD is still missing a sensitivity analysis and the disaggregation of suspended-solids data. The latter is relevant because soil erosion events are not only characterised by high water discharge due to overland flow but also by sediment mobilisation and high suspended-solids concentration. Thus, approximating the critical yield as product of overland flow per unit area (q_{fast} in chapter 3, q_f in Eq. 27) and suspended-solids concentration in overland flow (SSC_f) is more appropriate than as product of q_f and the observed mixed SSC (chapter 3). Like the "shortcut" disaggregation of total runoff q (Carl 2011), the conceptual modes of total suspended solids might be expressed as $SY_f(t)$, $SSC_f(t)$ (fast mode), $SY_t(t)$, $SSC_t(t)$ (transient mode), $SY_l(t)$, and $SSC_l(t)$ (slow mode) where t is the time (Eq. 27).

$$SY(t) = SY_f(t) + SY_t(t) + SY_l(t) = q(t) \cdot SSC(t) = q_f(t) \cdot SSC_f(t) + q_t(t) \cdot SSC_t(t) + q_l(t) \cdot SSC_l(t) \qquad \text{Eq. 27}$$

Third, the modelling of soil loss has to be revised to better match total or critical yields. This complex topic not only comprises improved USLE predictions but also identifying other sediment processes. There is ongoing research on the estimation of USLE factors and soil loss in Europe (P. Panagos, pers. comm.). For instance, Panagos et al. (2012) recently published a new USLE K factor map which is interpolated from measured soil properties instead of being approximated from the coarse information stored in the European Soil Database. In chapters 3 and 5, refining the regional estimation of the USLE R factor has been discussed as one step to overcome some limitations of the evaluated SDR approaches. Meusburger et al. (2012) currently proposed a new equation to estimate R factors in alpine areas (Switzerland). Similar efforts are needed in other regions in Europe, especially western Europe, because the intra-annual distribution of rainfall erosivity varies among regions and climate zones. Simply extrapolating regional approximations affects the distribution of calculated SDR and the relationships between catchment properties and these SDR.

Unfortunately, the disaggregation techniques tested in this dissertation thesis cannot discriminate between different soil erosion processes and sediment sources. Therefore, model adaptations should also aim at the question of sediment sources and sediment transport. How much of the estimated gross soil loss eventually contributes to the observed sediment yield at the catchment outlet? This in turn refers to the reliability of USLE predictions as well as to the importance of erosion processes not described by the USLE. Although the idea of a new alpine factor in the USLE to include avalanches and other snow-related processes (Konz Hohwieler 2010) is appealing, empirical models similar to the USLE are still not available for bank and gully erosion. Laubel et al. (2003) found an empirical relationship between bank erosion rate and bank properties, stream power and vegetation cover in Denmark. However, such data is not available for large-scale applications. For areas dominated by gully erosion in Spain, Martín-Fernández and Martínez-Núñez (2011) still had to rely on USLE outcomes to approximate soil loss rates. Three approaches are feasible to better the situation: i) evaluating sediment budgets in European river catchments, ii) evaluating the hysteresis effect during high water discharge (e.g., Lefrançois et al. 2007), and iii) applying empirical models for sediment yields instead of sediment delivery ratios, although findings in Spain have revealed that multiple regression models are outperformed by semi-quantitative factorial scoring models (Verstraeten et al. 2003; de Vente et al. 2005).

Fourth, the black box "sediment yield" (and any SDR derived from it) hides where sediments are deposited. As erosion models for river catchments are often used as management tools, separating in-stream and catchment retention allows better mitigating off-site effects of soil erosion (cf. Schäuble 2005; Kovacs et al. 2012). For instance, retention in streams and on flood plains is of great ecological and economic relevance and varies with sediment, runoff, vegetation, and stream characteristics (e.g., grain size, artificial barriers, macrophytes).

Unfortunately, sediment yields measured at the outlet are inadequate for independently validating the soil loss as well as catchment and in-stream retention of a river catchment. The spatial disaggregation of water discharge and sediment yield is, however, not straightforward and several disaggregation models have been proposed in hydrology (e.g., Tarboton et al. 1998; Prairie et al. 2007). A simple approximation of net in-stream retention is the difference between the sediment transport measured at nested monitoring gauges and reservoirs (Fuchs et al. 2012) – if the monitoring data is comparable and lateral fluxes are negligible. In case of ungauged tributaries between nested gauges, Eq. 26 can serve to close the gap in the sediment balance. For large-scale applications, however, the in-stream retention has to be regressed to stream and catchment properties.

Finally, the input data, algorithms and model assumptions in this dissertation thesis may be critically evaluated with more detailed data in regional studies. The approaches used in chapters 3 and 5 to estimate USLE parameters have been chosen after an extensive literature review. Nonetheless, the use of pan-European data necessarily limits the choice of appropriate algorithms and thus the evaluation of model uncertainties and sensitivities.

For instance, USLE C factors for crops are highly variable not only among but also within countries and regions. Consequently, the estimation of C factors is far more uncertain than just the difference between constant and crop-specific values in chapter 5. One should also be aware that the approach to use current land-use statistics in combination with crop-specific C factors is common but nonetheless questionable because USLE C factors are long-term means depending on crop rotation and rainfall distribution. Regression models between recent land use and measured C factors as used in chapter 3 (Auerswald 2002) are methodologically sound yet data demanding. They are still not available for large parts of Europe.

The estimation of the USLE L factors and delineation of water and sediment fluxes are also limited by the low data resolution and artefacts in DEM. Several algorithms have been developed to obtain L factors from DEM and (partly) ancillary data (as reviewed in Schäuble 1999). If high-resolution DEM and appropriate ancillary data are available, sophisticated algorithms should be compared to those applied in the previous chapters (e.g., Gisler et al. 2010).

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Appendix

Danksagung

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Eidesstattliche Erklärung

Hiermit versichere ich, dass ich die vorliegende Dissertation selbstständig und ohne unerlaubte Hilfe angefertigt habe. Ich versichere außerdem, dass ich die Dissertation an keiner anderen Universität eingereicht habe und keinen Doktorgrad im Fach Geographie besitze. Die dem Promotionsverfahren zugrunde liegende und im Amtlichen Mitteilungsblatt Nr. 34/2006 veröffentlichte Promotionsordnung ist mir bekannt.

Berlin,

Andreas Gericke