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Decadal modelling of rainfall–runoff erosivity in the Euro-Mediterranean region using extreme precipitation indices

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ABSTRACT

This paper presents and assesses the Decadal Rainfall Erosive Multiscale Model (DREMM), in which extreme precipitation data (95th percentile) are used to estimate decadal-scale rainfall–runoff erosivity values compatible with the Universal Soil Loss Equation and its revision – (R)USLE. The test area was a large region including central Europe and Mediterranean countries, in which 111 decades from 88 weather stations (ranging from about 3 to 1680 m above sea level) with rain and (R)USLE rainfall–runoff erosivity data were available. The construction of the model is simplified to a location-explicit term and to the understanding that the most erosive rainfalls are those recorded during the summertime and the beginning of autumn (May–October). These precipitation events are suitable for use in spatial and temporal climate variability studies on decadal time-scales. Once parameterized to capture decadal rainfall–runoff erosivity variability over central Europe and the Mediterranean, the DREMM was run to produce spatial patterns of rainfall–runoff erosivity in Germany and Bulgaria, compared to maps from the (R)USLE approach for 1961–1990. Implications for rainfall–runoff erosivity modelling were discussed concluding that a limited number of parameters may be sufficient to represent decadal rainfall–runoff erosivity for the central European and Mediterranean region. A transferable approach is recommended by employing the DREMM to assess the spatial impact of rainfall extremes. Its use to characterize the long-term dynamics of rainfall–runoff erosivity may also be of value in climate investigation.

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1. Introduction

Intense and extreme rainstorms are a prominent environmental threat facing humankind, especially affecting the agricultural sector. In particular, seasonal distribution of rainfall erosivity indicates that storms-erosivity are much more severe during summer months (Wischmeier and Smith, 1978; Auerswald et al., 2009), while the component of runoff of the erosivity factor is more balanced or major in fall season, especially in late autumn (Foster et al., 1977; Hrissanthou, 2005). Rainstorms' interannual variability is thus an important part of the climate signal to detect potential changes in the power of rainfall and extreme precipitation, which are events of great importance for society (Sauerborn et al., 1999; Easterling et al., 2000; D'Odorico et al., 2001; Nearing et al., 2004; Schütt et al., 2007; Verheijen et al., 2009).

This is especially so for the Central Europe and Mediterranean regions, focus of this paper, where human pressure and erratic rainfall patterns with marked interannual variability expose landforms to exacerbated, damaging hydrological processes (Sánchez et al., 2004; van Rompaey et al., 2005), and also to spur the emergence of new hazards, such as coastal and urban flooding (Papathoma and Dominey-Howes, 2003; Barroca et al., 2006).

Apart from changes in the mean, changes in the variability and the frequency of extremes of climate variables are of great importance for the modelling purposes (Katz and Brown, 1992; Schär et al., 2004; Schaeffer et al., 2005). However, variations in extreme precipitations have been studied less intensely than changes in means or totals, especially at global and continental scales (Kenyon and Hegerl, 2010).

In the European and Mediterranean area, efforts have been made in the last decades to reconstruct hydrological regimes from alluvial sequences (Poesen and Hooke, 1997; Bintliff, 2002) and to report trends in extreme precipitation indices (Alpert et al., 2002; Lionello et al., 2006). Some of the most important advances are those by Gaume et al. (2009) and Llasat et al. (2010) concerning the European projects Hydrate (<http://www.hydrate.tesaf.unipd.it>) and FLASH (<http://flash-eu>).

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tau.ac.il), respectively. These projects approached, for the first time, the creation of a homogeneous catalogue of multiple damaging hydrological events across regions in Europe and Mediterranean. However, there is still a lack of summary indicators for characterization of extreme hydrological events over long periods.

In this study, we refer to the rainfall–runoff erosivity index of the Universal Soil Loss Equation (USLE, [Wischmeier and Smith, 1978](#)). Without setting any of the environmental thresholds (e.g. floods and landslides) generally used to support weather warnings and forecasting, the rainfall–runoff erosivity index can be considered a suitable climatic indicator of the impacts of rainstorms across European and Mediterranean lands, where extreme rainfalls are dominated by a wide range of rainy events types (e.g. [Evrard et al., 2010](#)). Relationships for rainfall–runoff erosivity estimation remain however difficult to be developed on homogeneous bases of data over broad areas where spatial–time variability of climatic variables is high. Also in areas where sufficiently and detailed long time series are available, the number of locations is generally insufficient to extrapolate with confidence rainfall–runoff erosivity patterns for both spatial and temporal scales (after [Davison et al., 2005](#)). For this, attempts have been made to assess rainfall–runoff erosivity at large spatial scales. At global scale, for instance, [Yang et al. \(2003\)](#) identified alternate power and simple polynomial relationships between annual values of rainfall–runoff erosivity and precipitation.

[Diodato and Bellocchi \(2010a\)](#) approached the regional scale for multidecadal estimation of rainfall–runoff erosivity over Mediterranean. The European and Mediterranean climate shows a complex pattern of spatial and annual variability, which is exacerbated by the intrinsic unpredictability of interannual rainfall–runoff erosivity from rainfall amount only ([Diodato and Bellocchi, 2007, 2010a](#)). Climatic events such as droughts and storms are sometimes clustered into longer-term groups. This means that their impacts may not vary simply in relation to their size, thus complicating any simple magnitude and frequency relationship ([Diodato and Bellocchi, 2010b](#)).

Climatic variability is recognized as one key cause of these clusterings. The global mean precipitation response to climate change can vary significantly on 20-year timescales both as a result of internal climate variability and changes in radiative forcing ([Previdi and Liepert, 2008; Liepert and Previdi, 2009](#)). The implication of this is that recently observed rainfall changes may be a poor indicator of changes that occur in hydro-climatological processes (after [Previdi, 2010](#)). Improving estimates of heavy precipitation with the use of the fractional contribution of very wet days to total precipitation from the probability distribution of daily rainfall, as in European Climate Assessment Project ([Zolina et al., 2009](#)) can be useful to elicit information involved in damaging extreme hydrological processes. However, it is difficult to establish the link between these rainfall–indices and signals driving the rainfall–runoff erosivity over multidecadal spatio-temporal scales. A critical part of this link involves the evaluation of the spatio-temporal invariant of the model performance and scaling ([Diodato and Bellocchi, 2009](#)), with the difficulty in establishing whether a model is capable of correctly reproducing various statistics of the historical extreme rainfall climate signal (after [Westra, 2011](#)).

Rain rates and abstraction of extreme rainfalls over decadal scales could be a more suitable indicator to detect climatological forcing and to address hydrological patterns over both spatial and temporal scales. Following this idea, a novel rainfall–runoff erosivity model was developed and evaluated, linking the rainfall–runoff erosivity factor to readily available rainfall data to estimate decadal rates of rainfall–runoff erosivity (compatible with the USLE and its revision, RUSLE, by [Renard et al., 1997](#)) over the Euro-Mediterranean area.

The geographical representation chosen at a scale corresponding to central Europe and Mediterranean (intermediate on hemispheric

and local scales) guarantees distinct properties of the sub-continental scale while preserving parsimoniousness of the rainfall–runoff erosivity model. In particular, the Decadal Rainfall Erosive Multiscale Model (DREMM) developed in this study incorporates few of relevant variables such as monthly precipitation and geographic data that are able to capture climatic variability at multiple spatial scales and are easily available nowadays via, for instance, web resources.

2. Materials and methods

2.1. Study-area and data sources

The Central Europe and Mediterranean study-area (referred to as CEM hereafter) extends over about 5,000,000 km², centred across the Alps. The mapping of stations used ([Fig. 1](#)) roughly ranges from about 34° (Ain Boukellal, Morocco) to 54° (Suwalki, Poland) latitude north, and from about 8° west (Rio Leça, Portugal) to 35° east (Develi, Turkey) longitude. Elevations above sea level range from about 3 m of Pátrai (Greece) to 1680 m of La Molina (Spain), respectively. Annual rainfall–runoff erosivity values calculated with the (R)USLE – or with simplified methods of estimating the R-factor in (R)USLE – for different sites of central and Mediterranean Europe, North Africa and Middle East were derived from literature. In total, 111 data-points were available for this study, from 88 stations; multiple decades were available for six stations ([Appendix A](#)).

Within the global circulation system, central and Mediterranean Europe, North Africa and Middle East are located between the centres of action of the subtropical high pressure belt and the polar low pressure belt. Zonal and mixed circulations generally bring moist maritime airmasses from the tropic Atlantic into Europe, whereas meridional circulation can bring maritime polar airmasses or dry continental airmasses (polar or tropic source).

Numerous peaks and valleys characterize the CEM countries. This variegated morphology has important consequences on both sea and atmospheric circulations, which determine a non-uniform distribution of weather types ([Lionello et al., 2006; James, 2007](#)) and a large spectrum of hydrological events, found to vary considerably both over time and geographically ([Macklin et al., 2006](#)). A data analysis of extreme precipitation events in the CEM countries shows a distinct thermo-convection mode occurring in the afternoon over summer season (May–September), which takes about 3 h ([van Delden, 2001; Twrdosz, 2007](#)), and a cyclonic advection mode with precipitation duration of about 10 h ([Kottmeier and Kunz, 2002; Twrdosz, 2007](#)). Monthly rainfalls were derived from ECAD-European Climate Assessment & Dataset (ECAD, <http://eca.knmi.nl>), and the Global Historical Climatology Networks (GHCN, <http://www.ncdc.noaa.gov/gchcnm>) via Climate Explorer ([Van Oldenborgh et al., 2009](#)).

2.2. Multiscale model of rainfall–runoff erosivity

Multiple linear and nonlinear regression models can be an option to account for spatial and temporal heterogeneity of a given variable. However, it may not be immediately clear what order of polynomial should be used, and a wrong expansion may hide important local variations in the model form ([Charlton and Fotheringham, 2009](#)).

Multi-level modelling is an alternative approach (after [Goldstein, 1987](#)), which combines models for individual and spatially aggregated characteristics within the same overall model. The underlying idea of a multiscale model is that parameters may be estimated anywhere in time and space given a dependent variable and a set of one or more independent variables, measured at places whose location is known. This same idea stems from the principle of Geographically Weighted Regression (e.g. [Fotheringham et al., 2002](#)), a version of spatial regression, which generates parameters disaggregated by spatial units

Fig. 1. Location of central European and Mediterranean stations for use in decadal rainfall–runoff erosivity modelling. Circles and squares represent stations used for calibration and validation, respectively (as in Appendix A).

and whose typical expression is a linear combination of independent terms:

$$y_i(\mathbf{u}) = \beta_{0i}(\mathbf{u}) + \beta_{1i}(\mathbf{u})x_{1i} + \beta_{2i}(\mathbf{u})x_{2i} + \dots + \beta_{mi}(\mathbf{u})x_{mi} \quad (1)$$

The notation $\beta_{0i}(\mathbf{u}) + \dots + \beta_{mi}(\mathbf{u})$ indicates that each parameter β describes a relationship around location $\mathbf{u}$ (vector $\mathbf{u}$ representing the coordinates of location) and is specific to that location. Prediction can be made for the dependent variable (y_i) if measurements for the independent variables (x_i) are also available at location $\mathbf{u}$.

To assess the amount of multi-year erosive events, series of monthly precipitation data may provide adequate time resolution for multiscale modelling purposes. Geographic elements were addressed in previous contributions (van der Knijff et al., 2000; Diodato and Bellocchi, 2007, 2010a) to estimate rainfall–runoff erosivity in the CEM zones.

In this study, the same principle was expanded to achieve a satisfactory solution in which monthly rainfall quantiles and the geographical control are modelled together to account for temporal and spatial dependence of rainfall–runoff erosivity. It is assumed that decadal mean annual rainfall–runoff erosivity values is a function of a quantile of the monthly precipitation distribution over the same decade, and that the parameters of this quantile vary with the raingauge location. Based on this understanding, summertime extreme rains are captured by percentile statistics across the months of May, June, July, August and September. Application of the general quantile regression technique has already yielded relationships between the rainfall–runoff erosivity factor and precipitation (e.g. Pelacani et al., 2008; Diodato et al., 2011).

The complex hydrological-erosivity system was deconstructed while preserving its representation for application-specific operability.

In particular, the conceptual model was resolved into a non-linear equation with parsimonious structure:

$$R_{\text{DREMM}} = k \cdot (P_{\text{prc95}(M-S)})^\eta + \alpha \cdot \sqrt{P_{(\text{max})\text{Oct}}} \quad (2)$$

where R_{DREMM} ($\text{MJ mm ha}^{-1} \text{h}^{-1} \text{y}^{-1}$) is the estimated decadal mean rainfall–runoff erosivity; $P_{\text{prc95}(M-S)}$ (mm) is the 95th percentile of the monthly rainfall from May (M) to September (S) over each decade; $P_{(\text{max})\text{Oct}}$ (mm) is the maximum monthly rainfall in October over the decade; k ($\text{MJ mm}^\eta \text{ha}^{-1} \text{h}^{-1} \text{y}^{-1}$) and α ($\text{MJ mm}^{0.5} \text{ha}^{-1} \text{h}^{-1} \text{y}^{-1}$) is a scale parameter; η is a shape term depending on the geographic location, in the following form:

$$\eta = \beta + \gamma \cdot \text{LAT} + \delta \cdot \text{LONG} \quad (3)$$

where β is a shift parameter (a virtual value of η at the point at which the equator and the prime meridian intersect), γ ($^\circ^{-1}$) and δ ($^\circ^{-1}$) are scale parameters.

The parameters of Eqs. (2) and (3) were optimized by a co-iterative calibration process against mean decadal (R)USLE rainfall–runoff erosivity values. The best fit was obtained of a regression equation $Y = a + b \cdot X$, where Y = model estimates and X = actual data, according to the following criteria:

$$\begin{cases} r^2 = \max \\ NS = \max \\ |b-1| = \min \end{cases} \quad (4)$$

First, the goodness-of-fit (r^2 , optimum = 1) and the Nash–Sutcliffe (NS, optimum = 1) efficiency index (Nash and Sutcliffe, 1970) of the

Table 1

Parameter values estimated from the entire dataset of stations and decades for the DREMM (Eq. (2)) and the exponent η (Eq. (3)).

Parameter	DREMM	η
α (MJ mm ^{0.5} ha ⁻¹ h ⁻¹ y ⁻¹)	24	–
β	–	2.45950
γ (°–1)	–	–0.02266
δ (°–1)	–	–0.00477

linear function were maximized, and then the unit slope (b) of the straight line that would minimize the bias was approximated. As additional measures, the intercept of the regression equation (a , optimum=0) and the mean absolute error (MAE, optimum=0) were also calculated. The calibration process was performed over a subset of stations and decades, while 26 data (26 stations with one decade each, roughly 1/4 of the total number of data) were held back randomly for validation (Appendix A). In order to provide proof of temporal-scale invariance, model performance was also shown for the six stations for which multiple decades were available (29 data).

The scale parameter k to convert the first term of Eq. (2) – $P_{\text{prc95}(M-S)}^{\eta}$ – to MJ mm ha⁻¹ h⁻¹ y⁻¹ was initially set equal to one and, for reasons of parsimony as given in Grace (2004), not treated as a free parameter because the initial value resulted in a fit that satisfied the criteria outlined by Eq. (4). The process was assessed interactively using Microsoft® Office Excel 2003 with the support of statistics software STATGRAPHIC Online (<http://www.statgraphicsonline.com>).

Parameter η was also calibrated. Calibrated values of η are those optimized against data once parameters β , γ and δ of Eq. (2) are calibrated. In this case, 88 data were used for calibration, that is, only one decade per station was taken to ensure the same weight of each data-record (the median value was used for stations where multiple decades were available). Calibrated values of η were compared against calculated values produced by Eq. (3).

A correlation analysis between terrain attributes and rainfall–runoff erosivity data was performed by geostatistical wizard in ArcGIS-ESRI, release 9.3 (Johnston, 2004). Terrain attributes are the environmental variables most commonly used with the hydrological data, since topography is an important forcing factor for erosive rainfall patterns. However, the evaluation of the effect of topography on precipitation is not straightforward. Elevation, in particular, is among the most relevant terrain variables controlling the precipitation pattern. The scales at which orographic effects are observed are critical for a region, with resolutions commonly varying between 1 and 3 km (Diodato, 2006).

In our case, Digital Elevation Model (DEM) with 1-km resolution was selected as an appropriate scale to account for terrain and station point features. The same ESRI-GIS tool was employed to create at random validation sites for the DREMM (grey cells in Appendix A).

3. Results and discussion

In this section, comparison between (R)USLE-based ('actual') and DREMM (modelled) rainfall–runoff erosivity data are presented and discussed. The assumptions behind the modelling solution of Eq. (2) are also discussed (nature of the variable exponent η , window of months for estimation of decadal rainfall–runoff erosivity, co-evolution between erosivity and percentiles of rainfall, terrain characteristics). Rainfall–runoff erosivity patterns in nearby areas provided the basis for qualitative validation of the DREMM.

3.1. Model parameterization and evaluation

The optimized parameters of Eqs. (2) and (3) determined against (R)USLE rainfall–runoff erosivity data are given in Table 1.

Fig. 2a reports the calibration results for 85 data-points (as in Appendix A), where negligible departures of the data-points from the 1:1 line are observed. The model calibration was evaluated quantitatively based on the correlation and amount of residuals between the estimated and the actual data. The Nash–Sutcliffe (NS) efficiency index and the correlation coefficient (r), equal to 0.92 and 0.95 (Table 2), are satisfactory. This is also due to the ability of Eq. (3) to capture the variability of the shape term η , as assessed by comparing modelled to calibrated results (Fig. 2a₁; NS=0.90, r =0.93, Table 2). The estimated exponent values are greater than one, which means that a relatively small increase in the $P_{\text{prc95}(M-S)}$ values will produce a large change in decadal rainfall–runoff erosivity.

In order to assess the relevance of time scales, the performance of the DREMM was also examined by showing predicted against (R)USLE rainfall–runoff erosivity data for the stations where multiple decades were available (Fig. 2b; NS=0.83, r =0.92, Table 2). The mean absolute error equal to 100 MJ mm ha⁻¹ h⁻¹ y⁻¹ is similar to that obtained over the calibration dataset (MAE=111 MJ mm ha⁻¹ h⁻¹ y⁻¹, Table 2), thus indicating some stability of the model in Eq. (2) on both spatial and temporal scales (scale invariance).

Independence-of-errors due to the possible presence of significant serial autocorrelations among the residuals was also tested. Strong temporal dependence may in fact induce spurious relations (see Granger et al., 2001). The Durbin–Watson statistic (Durbin and Watson, 1950, 1951), which tests the residuals to determine if there is any significant correlation based on the order in which they occur in the time series,

Fig. 2. Model calibration: (a) scatterplot between modelled by Eq. (2) and actual rainfall–runoff erosivity (calibration against the dataset of 85 stations and decades, as in Appendix A); (b) parameterization as in (a) but assessment for the six stations with multiple decades only; (a₁) scatterplot between modelled and calibrated η term of Eq. (3) (calibration against the dataset of 88 stations, taking one decade per station). Rainfall–runoff erosivity is expressed in MJ mm ha⁻¹ h⁻¹ y⁻¹; η is dimensionless. In each graph, the black line is the interpolating line, the bold grey line denotes the 1:1 line, and grey curves are 0.99 confidence limits of the interpolating line. Frequency distributions are reported on both axes.

Table 2Performance and autocorrelation statistics, calculated on alternative datasets for the DREMM (Eq. (2)) and the exponent η (Eq. (3)).

Comparisons and datasets	Performance statistics				Autocorrelation statistics		
	Least-squares regression		Nash-Sutcliffe efficiency coefficient (NS)	Correlation coefficient (r)	Mean absolute error (MAE)	Lag-1 residual correlation	Durbin–Watson statistic
	Intercept (a)	Slope (b)					
1. Modelled (Eq. (2)) versus (R)USLE rainfall–runoff erosivity (calibration against the dataset of 85 stations and decades, Appendix A)	49	0.96	0.92	0.95	111 ^a	0.11	1.65 ($p = 0.05$)
2. Modelled (Eq. (3)) versus calibrated η (dataset as 1, with one decade per station)	−5.70E−03	1.00	0.90	0.93	0.03 ^b	−0.02	2.03 ($p = 0.52$)
3. Modelled (Eq. (2)) versus (R)USLE rainfall–runoff erosivity (dataset of stations with multiple decades; parameterization as in 1)	127	0.87	0.83	0.92	100 ^a	−0.32	2.61 ($p = 0.94$)
4. Modelled (Eq. (2)) versus (R)USLE rainfall–runoff erosivity (dataset with 26 stations, Appendix A; parameterization as in 1)	−2.07	0.99	0.92	0.96	123 ^a	−0.31	2.47 ($p = 0.86$)

^a MJ mm ha^{−1} h^{−1} y^{−1}.^b Unitless.

indicates that serial autocorrelation in the residuals of the three equations are not significant ($p \geq 0.05$).

Fig. 3 and Table 2 also show that the performance of the model (Eq. (2)) is still satisfactory when run against the independent set of 26 stations (as in Appendix A).

3.2. Modelling assumptions: temporal and spatial patterns

The assumptions that underly a model must be well understood and explicitly stated with reference to the conditions under which they are valid (Mulligan and Wainwright, 2004).

The DREMM in Eq. (2) represents rainfall–runoff erosivity through a power-law function involving a variable exponent. The scale parameter α is a conversion factor that can be conveniently assumed constant over time and space. The variable exponent is appealing because, not only does it provide a parsimonious description but, there is also a generic mechanism of the process. In general, geographic location is known to be an important input property to rainfall–runoff erosivity models because the location, and then the climate zone, accommodates a broad range of conditions related to the occurrence of abundant and intensive precipitations (e.g. Diodato and Bellocchi, 2010a). In Eq. (3), we assumed that the variable exponent may continuously vary with latitude and longitude (Fig. 4), as a shape term to modulate the

percentile statistic that pulls out seasonal erosive rainfall between May and September (the warm season). Latitude appears to play a prominent role in the variability of η , with increasing values from northern latitudes (just above one) to more southern regions of Europe (up to about 1.8). Based on longitude, the magnitude of variability is much lower (up to about 20%). This corresponds to the knowledge that high-intensity storms mainly affect areas from the subtropics to the lower temperate zones (e.g. Mariotti et al., 2005). It is also important to appreciate that the shape power-law exponent of Eq. (3) is a mechanism that serves to either attenuate or enhance rainfall–runoff erosivity depending on site-specific climate conditions. In the warm season, in fact, daily and sub-daily storms can be accompanied by high rain variability and intensity, thus releasing a large amount of energy through sparse phenomena, with rain-peaks generally concentrated in few hours (Diodato, 2004; Barrera et al., 2006; Twrdosz, 2007). This was also supported by the regime of within-year floods (Fig. 5b), including floods caused by heavy rains (flash floods).

An important issue at the stage of model development has been the choice of the window of months for estimation of decadal rainfall–runoff erosivity. In general, intense rainstorms fall between the end of spring and start of autumn. In the Northern hemisphere, the flood regimes that correspond to ice-jam floods over December–January, snowmelt floods in November and over March–April, and a mix of

Fig. 3. Model validation: scatterplot between modelled by Eq. (2) and actual rainfall–runoff erosivity for the dataset of 26 stations as in Appendix A. The black line is the interpolating line, the bold grey line denotes the 1:1 line, and grey curves are 0.99 confidence limits of the interpolating line.

Fig. 4. Variation of the η exponent (Eq. (3)) over the range of latitudes (LAT, °) and longitudes (LONG, °) covered by the study-area.

Fig. 5. Precipitation rates and hydrological regimes in Europe: (a) seasonal regime of rain rate; (b) monthly frequency of floods and flash-floods. The rain rates are arranged by NCEP/NCAR reanalysis at NOAA/ESRL Physical Science Division over 1947–2007 (Kalnay et al., 1996). The frequencies of floods are for Europe (27 countries) over 1947–2007, extracted from Guha-Sapir et al. (2010). The frequencies of flash-floods are an elaboration over 1947–2006 from surveys at different European countries: Austria, France, Italy, Romania and Slovakia (Gaume et al., 2009); Bulgaria (Bocheva et al., 2007); Germany (van Delden, 2001); Spain (Barrera et al., 2006; Llasat et al., 2007; Gaume et al., 2009). The most powerful rain events in two range-bound periods of six (May to October, graph a) and five (June to October, graph b) months are expected to drive together major hydrological extreme events and generate multiple damaging hydrological events.

both in February (Radziejewski, 2011) do not contribute to rainstorm erosivity. However, also in the warm season the climate varies greatly in different areas of the CEM region. This is why boundary months at either end of this period may be critical for rainfall–runoff erosivity modelling. It was observed, for instance, that the precipitations collected in May may not be an important driver of rainfall–runoff erosivity prediction in northern and eastern Europe, while they are useful inputs in western Europe and Mediterranean (data not shown). In the month of October, rain events present more homogeneous features most of the time, especially in western Europe (van Delden, 2001). Using the reanalysis data from ERA-40 (<http://www.ecmwf.int/research/era/do/get/era-40>) for the period 1971–2000, Romero et al. (2007) described severe convective storms occurring between June and October in Europe. For Europe, the occurrence of severe thunderstorms is indeed intimately associated with the convective environmental conditions (Brooks et al., 2007). An analysis of precipitation rates and hydrological regimes also reveals that they are not uniform for different months (Fig. 5). The frequencies of flood and flash-flood events in Fig. 5b are from data-sparse basins in Europe. It is shown that the precipitation in May likely leads to a different regime of hydrological extremes compared to June–October, when flash-floods are more frequent (Fig. 5b).

This means that the period June–October delineates a key time window in estimating rainfall–runoff erosivity in Europe but, on the other hand, precipitations rate at the 75th percentile (interpolated over Europe on 2.5° latitude–longitude grid resolution) are more frequent in May than in late autumn (Fig. 5a). It is in this respect that we have taken October as a transition month. Its contribution to rainfall–runoff erosivity is less pronounced as the European continent may have rain rates less intense in this month than in prior months (Fig. 5a). Moreover, in the CEM region, the rains and storms become more extended and with longer duration with the continuing of the autumn season, thus exhibiting relevant geomorphological effectiveness towards more runoff–erosivity than the splash–erosive fraction (Diodato, 2006). This is why October rainfall was introduced as an additive component to the DREMM, i.e. $P_{(\max)Oct}$ in Eq. (2), held out of the percentile calculation based on May to September precipitations. For October, the maximum precipitation amount over the decade ($P_{(\max)Oct}$) returns a non-linear effect (square root), directly converted into a rainfall–runoff erosivity fraction via the scale parameter α . As we intend to show, satisfactory model performance can be obtained from some reasonable and physically sound assumptions (as illustrated above). Other modelling solutions were

explored, essentially by changing the seasonal window over which the statistical percentile works. For instance, a simple solution was to calculate the 95th percentile during the window from May to September; another solution mirrored the final DREMM structure, but considering the window from June to October for the percentile calculation, and taking the square root of the May maximum precipitation as additive component. These and other avenues of model development were dropped because they performed well in some areas but not across the whole of the CEM. The model in Eq. (1) had the best fit, was the most parsimonious, and was consistent with the literature. However, the seasonal window over which the statistical percentile works remains a critical assumption and one that requires review in the future to ensure the reliability of DREMM estimates at sites where pluviometric series are missing (for punctual estimation, interpolation or validation purposes).

In an effort to provide rigor to the methodology, along with the ability to represent time-varying rainfall–runoff erosivity values, we compared extended summer (May–September) erosivity density (ED , ratio between annual values of rainfall–runoff erosivity and rainfall total) to the 95th percentile ratio (PR between the same season and the remaining months (the value of rainfall below which 95% of the observed rain amounts fall over May–September divided by the same amount over the other months), calculated for Ukkel (Belgium), Naples (Italy) and Danube river basin (near Bratislava, Slovakia) for 1898–2006 (Fig. 6). From visual inspection, Fig. 6 shows that there is moderate co-evolution between ED and PR for both the towns of Ukkel, Belgium (graph a, correlation coefficient equal to 0.58) and Naples, Italy (graph b, correlation coefficient equal to 0.46), and for the Danube river basin (graph c, correlation coefficient equal to 0.86). This suggests that the DREMM (as developed and parameterized in this study) is sufficiently robust and may be liable to a spatial generalization for different climates, either dominated by Atlantic influences (Ukkel), continental regime (Danube river) or Mediterranean character (Naples).

Assumptions about representative terrain characteristics are regarded as playing a central role in hydroclimatology. Also, when these features are not explicitly included in the models, model outcomes and terrain co-variables should be comparable in the sense of having similar distribution. In our case, we have compared the elevation distribution of raingauge stations with the terrain elevation extracted from a digital elevation model (DEM) of 1-km resolution. Fig. 7a shows that stations around hilly terrains (at about 400 to 800 m a.s.l.) tend to be more biased, compared to low- and high-altitude sites,

Fig. 6. Moving-window average on previous 10 years of annual erosivity density (*ED*, grey curve), and 95th percentile ratio of monthly rainfall (*PR*) between the extended summer season (May–September) and the remaining months (bold curve) calculated at three European sites for 1898–2006: (a) Ukkel, Belgium; (b) Naples, Italy; (c) Danube river basin near Bratislava, Slovakia. Rainfall–runoff erosivity data were arranged from [Chirico et al. \(2010\)](#), [Verstraeten et al. \(2006\)](#) and [Pekarova et al. \(2008\)](#), respectively.

because the set of station data from mid-mountain locations is not entirely representative of the distribution of the DEM orography. In addition, in order to elucidate the relationships between the DREMM and terrain features, cross-covariance functions were examined for discovering sub-regional characteristics of spatial correlation between rainfall–runoff erosivity and elevation. Bit cells through which spatial shifts in correlation between stations altitude and both actual ([Fig. 7b₁](#)) and predicted ([Fig. 7b₂](#)) rainfall–runoff erosivity values are given. The cross-covariance surfaces reflect a similar pattern (although less pronounced with the DREMM), thus indicating that the DREMM may serve as a suitable model to generate spatial patterns of rainfall–runoff erosivity response consistent with the terrain complexity observed at the regional scale if the inter-station distance is less than 200 km. For larger distances, the correlation between stations is not enough to be useful about the accuracy of patterns of response over large areas.

3.3. Evaluation of spatial patterns

We have applied the DREMM (parameterized against the dataset of 85 stations and decades as in [Appendix A](#)) to compare spatial patterns of rainfall–runoff erosivity ([Fig. 8](#)) and maps compatible with the (R)USLE approach for Germany ([Hennings, 2003](#)) and Bulgaria ([Rousseva and Stefanova, 2006](#)). Overall, the similar pattern generally observed between estimated and (R)USLE-calculated data suggests the possibility for the DREMM to be used for the construction of erosive maps at regional scale. Discrepancies between DREMM and (R)USLE

data are indeed few and isolated. They may depend on the large scale on which the DREMM was developed with respect to the local scale used to compute the (R)USLE data. Moreover, some biases may have been introduced by the GPCC-V5 rainfall dataset, which includes smoothed gridded values and not the original data points.

4. Conclusions

Parsimonious hydro-climatological models hold potentials to simulate the combined effects of rainfall input patterns in the absence of precipitation records at short time intervals (e.g. daily to sub-hourly) and over large areas (e.g. regional to continental). The present study offers a parsimonious model of rainfall–runoff erosivity, the DREMM, which was evaluated and effectively used to assess the distribution of decadal mean rainfall–runoff erosivity over the CEM. It shows that the model holds potential to simulate decadal rainfall–runoff variability on sub-continental spatial scales. Its results need to be interpreted with caution to analyse erosivity in detail at the sub-regional versus local scales, where including a site-specific elevation term (with or without re-parameterization) could potentially improve the performance of the model (e.g. [Meusburger et al., 2011](#)). A different approach (e.g. [Martí and Gasque, 2011](#)) could be adding ancillary (local) pluviometric stations across the grid points (based on international datasets) in cases where the effect of elevation on rainfall–runoff erosivity is not reflected by a clear relationship. The limited number of stations used in this study also raises the need for refinement of the model and wider validation work. However, the satisfactory performance, which

Fig. 7. Model outcomes and terrain co-variables: (a) the (cumulative) frequency functions of the elevation pattern (1×1 km DEM, bold black curve), and altitude of the 88 rain gauge stations of [Fig. 1](#) (thin black curve); (b₁) spatial cross-covariance (CC) between the actual rainfall–runoff erosivity and stations altitude; (b₂): spatial cross-covariance (CC) between the DREMM rainfall–runoff erosivity and stations altitude. Spatial cross-covariance is depicted by bit-colour maps; the circle overimposed in b₁ and b₂ represent the range of 200 km, elected from cross-covariance models to indicate the spatial scale (distances to which the attributes are more similar) and for inferring the spatial co-dependence between rainfall–runoff erosivity and altitude of stations.

Fig. 8. Rainfall-runoff erosivity spatial patterns (averaged over the period 1961–1990): (a) Germany, by Hennings (2003); (a₁) Germany, by the DREMM; (b) Bulgaria, by Rousseva and Stefanova (2006); (b₁) Bulgaria, by the DREMM. Baseline rainfall gridded data points for spatialization of DREMM values (a) were extracted via the web-GIS Climate Explorer (Van Oldenborgh et al., 2009), based on GPCC-V5 rain gauge precipitation dataset (http://climexp.knmi.nl/select.cgi?someone@somewhere+gpcc_05). Rainfall-runoff erosivity values in Fig. 8b₁ are in 10⁻¹ MJ mm ha⁻¹ h⁻¹ y⁻¹.

we have been able to achieve with the DREMM, is the result of sound assumptions and proper modelling of terms that constitute the essential of large-scale precipitation regime. In dealing with such a large and heterogeneous region, which includes stations from North Africa to the Atlantic–Baltic sector and from Iberian Peninsula to Middle East, there are constraints on the seasonal variability of extreme rainfalls that we can explain with just a small number of inputs. This confirms that regional-scale (sub-continental) modes of variability are important and extreme rainfall is not just controlled by local processes (Haylock and Goodess, 2004). Future work should focus on delineating the extents of these promising findings (e.g. at other portions of the Northern hemisphere) as well as clarifying the mechanisms for shaping broad-scale variations in extreme storms and rainfall erosivity. In addition, the DREMM approach lays the foundation for the reconstruction of

historical hydroclimatic conditions, when detailed precipitation records are unavailable.

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Appendix A. Station sites, climate summaries, data availability. The total number of stations is 88, six of which have multiple decades. Grey cells indicate 26 stations and decades used for validation of Eq. (2)

Country	Station	Latitude (degree)	Longitude (degree)	Elevation (m)	Actual rainfall- runoff erosivity (MJ mm ha ⁻¹ h ⁻¹ y ⁻¹)	Period (years)	Sources
Algeria	Batna	35.60	6.20	1000	813	1969–2001	Khentouche (2005)
Austria	Retz	48.75	16.00	243	550	1971–1980	Strauss et al. (1995)
	Weiz	47.00	15.63	480	1380	1976–1990	Strauss et al. (1995)
Belgium	Ukkel	50.80	4.40	104	815	1898–1907	Verstraeten et al. (2006)
	Ukkel	50.80	4.40	104	894	1908–1917	Verstraeten et al. (2006)
	Ukkel	50.80	4.40	104	844	1918–1927	Verstraeten et al. (2006)
	Ukkel	50.80	4.40	104	820	1928–1937	Verstraeten et al. (2006)
	Ukkel	50.80	4.40	104	768	1938–1947	Verstraeten et al. (2006)
	Ukkel	50.80	4.40	104	807	1948–1957	Verstraeten et al. (2006)
	Ukkel	50.80	4.40	104	965	1958–1967	Verstraeten et al. (2006)
	Ukkel	50.80	4.40	104	841	1968–1977	Verstraeten et al. (2006)
	Ukkel	50.80	4.40	104	824	1978–1987	Verstraeten et al. (2006)
	Ukkel	50.80	4.40	104	985	1988–1997	Verstraeten et al. (2006)
Bosnia-Erzegovina	Sarajevo	43.80	18.20	511	1500	1953–1980	Arranged from Petković et al. (1999)
Bulgaria	Kolarovgrad	43.30	26.90	198	911	1960–1979	Rousseva and Stefanova (2006)
	Pleven	43.40	24.60	75	1000	1960–1979	Rousseva and Stefanova (2006)
	Sofia	42.70	23.40	550	1170	1960–1979	Rousseva and Stefanova (2006)
Czech Republic	Brno River	49.20	16.70	360	520	1980–2000	Krasa et al. (2005)
	Vrchlice Reservoir	50.00	15.30	254	480	1971–2000	Krasa et al. (2005)
	Liberec	50.70	15.00	374	760	1961–2000	Krasa et al. (2007)
France	Bennwihr	48.15	7.32	330	540	1966–1994	Strauss et al. (1997)
	Gap	44.57	6.07	773	1650	1951–1970	Pihan (1978)
	Horbouir-Wihr	48.10	7.40	190	670	1968–1994	Strauss et al. (1997)
	Hunspach	48.95	7.95	150	720	1976–1994	Strauss et al. (1997)
	Orleans	48.00	1.80	125	680	1951–1970	Pihan (1978)
	Rouen	49.40	1.20	879	879	1959–1988	Arranged from Bollinne (1979)
	Stenay	49.50	5.20	180	1455	1950–2000	Ward et al. (2009)
	Toulouse	43.60	1.44	187	1275	1951–1970	Pihan (1978)
	Valence	44.95	4.90	106	1700	1951–1970	Pihan (1978)
Germany	Eichstrutt	48.88	11.18	340	700	1961–1980	Schweikle et al. (1985)
	Freudenstadt	48.50	8.40	728	1600	1961–1980	Schweikle et al. (1985)
	Hamburg	53.60	10.00	16	550	1961–1990	Wurbs and Steininger (2011)
	Kluppenek	48.10	8.75	980	960	1951–1970	Schweikle et al. (1985)
	Leipzig	51.30	12.40	100	667	1961–1970	Arranged from Hennings (2003)
	Leipzig	51.30	12.40	100	458	1971–1980	Arranged from Hennings (2003)
	Leipzig	51.30	12.40	100	369	1981–1990	Arranged from Hennings (2003)
	Leipzig	51.30	12.40	100	655	1991–2000	Arranged from Hennings (2003)
	Leipzig	51.30	12.40	100	771	2001–2010	Arranged from Hennings (2003)
	Mannheim	49.50	8.50	97	710	1951–1980	Schweikle et al. (1985)
	München	48.10	11.56	518	1394	1955–1974	Rogler and Schwertmann (1981)
	Villingen	48.10	8.50	800	800	1951–1970	Strauss et al. (1997)
Greece	Patrai	38.20	29.70	3	558	1961–1990	Arranged from Hrisanthou and Piliotis (1995)
	Venetikos River	40.09	21.15	900	812	1974–1983	Panagoulia et al. (2004)
Hungary	Gyöngyösoroszi	47.80	19.90	881	853	1961–1990	Tamás and Kovács (2003)
	Gyor	47.70	17.70	129	670	1961–1980	Thyll (1992)
	Nagykanizsa	46.50	17.00	100	880	1961–1980	Thyll (1992)
	Sósi Creek River	47.70	19.50	150	685	1961–1990	Centeri et al. (2008)
	Turkeve	47.40	21.00	94	588	1961–1980	Thyll (1992)
Italy	Borgo S.Lorenzo	41.46	12.90	12	1503	1998–2004	Diodato and Bellocchi (2010a)
	Campochiaro	41.50	14.52	502	1780	1994–2003	Diodato and Bellocchi (2010a)
	Carpeneto	44.70	8.60	230	1432	1994–2003	Diodato and Bellocchi (2010a)
	Cesena	44.00	12.20	44	1165	1951–1960	Biagi et al. (1995)
	Cesena	44.00	12.20	44	1343	1961–1970	Biagi et al. (1995)
	Cesena	44.00	12.20	44	1401	1971–1980	Biagi et al. (1995)
	Enna	37.57	14.27	940	1324	1951–1993	D'Asaro et al. (2007)

(continued on next page)

Appendix A (continued)

Country	Station	Latitude (degree)	Longitude (degree)	Elevation (m)	Actual rainfall- runoff erosivity (MJ mm ha ⁻¹ h ⁻¹ y ⁻¹)	Period (years)	Sources
	Monsampolo	42.88	13.80	230	1300	1994–2003	Diodato and Bellocchi (2010a)
	Monteombraro	44.40	11.20	727	1530	1956–1985	Calzolari et al. (2001)
	Palo del Colle	41.00	16.70	177	808	1994–2003	Diodato and Bellocchi (2010a)
	Piano del Leone	37.68	13.60	250	1073	1951–1970	D'Asaro et al. (2007)
	Pontecagnano	40.62	14.87	29	1478	1994–2003	Diodato and Bellocchi (2010a)
	Ramacca	37.38	14.70	270	1589	1951–1970	D'Asaro et al. (2007)
	Razzuolo	44.00	11.50	650	2223	1981–2000	Borselli et al. (2004)
	San Pietro	37.57	15.00	330	1032	1995–2004	Diodato and Bellocchi (2010a)
	Sciacca	37.50	13.08	60	674	1951–1970	D'Asaro et al. (2007)
	Siracusa	37.08	15.28	17	1054	1951–1970	D'Asaro et al. (2007)
	Volterra	43.40	10.80	151	852	1964–1972	Bazzoffi and Pellegrini (1991)
	Volterra	43.40	10.80	151	1034	1973–1981	Bazzoffi and Pellegrini (1991)
	Volterra	43.40	10.80	151	1313	1982–1990	Bazzoffi and Pellegrini (1991)
Morocco	Ain Boukellal	34.16	-4.12	500	957	1977–1994	Tribak et al. (2009)
Netherlands	de Bilt	52.00	5.13	5	799	1955–1974	Bergsma (1980)
Poland	Elblag	54.07	19.40	40	600	1960–1988	Baryla (2006)
	Limanowa	49.70	20.40	400	968	1960–1988	Banasik and Górski (2000)
	Olszanka	52.15	22.70	120	613	1951–1990	Olszty and Kowalskiego (2005)
	Otwork	52.10	21.40	100	580	1984–2002	Arranged from Baryla (2004)
	Puczniew	51.78	19.00	110	285	1983–1992	Baryla (2006)
	Puczniew	51.78	19.00	110	483	1993–2002	Baryla (2006)
	Sandomierz	50.70	21.70	200	660	1960–1988	Banasik and Górski (2000)
	Swinoujście	53.90	14.30	5	370	1961–1980	Banasik and Górski (2000)
Portugal	Suwalki	54.00	22.90	170	426	1960–1988	Banasik and Górski (2000)
	Rio Leça (Porto)	41.20	-8.70	77	2993	1983–1994	Petan et al. (2009)
Romania	Lagos	37.08	-8.67	35	1117	1971–1997	de Santos Loureiro and de Azevedo Couthino (2001)
	Valea Calugareasca	44.97	26.15	140	871	2002–2007	Cardei et al. (2009)
Serbia	Zabala River	45.88	26.18	600	913	1961–1990	Patriche et al. (2006)
	Snagovo	44.70	19.20	146	1391	1973–1990	Svetlana et al. (2006)
Slovakia	Handlova	48.73	18.75	420	630	1971–1990	Šúri et al. (2002)
	Oravska	49.50	19.50	500	789	1971–1980	Maderková and Antal (2008)
Slovenia	Solcava	46.40	14.70	658	2258	1991–2010	Mikoš et al. (2006)
	Murska Sobota	46.70	16.10	190	1200	1991–2006	Ceglar et al. (2008)
	Ratece	46.50	13.70	900	1788	1991–2000	Ceglar et al. (2008)
Spain	Biescas	42.63	-0.32	875	2076	1971–1992	Renschler et al. (1999)
	Canelles	42.00	0.30	700	1517	1937–1946	Arranged from López-Vicente et al. (2007)
	Canelles	42.00	0.30	700	1043	1947–1956	Arranged from López-Vicente et al. (2007)
	Canelles	42.00	0.30	700	1675	1967–1976	Arranged from López-Vicente et al. (2007)
	Canelles	42.00	0.30	700	1034	1977–1986	Arranged from López-Vicente et al. (2007)
	Canelles	42.00	0.30	700	1526	1987–1996	Arranged from López-Vicente et al. (2007)
	Canelles	42.00	0.30	700	994	1997–2006	López-Vicente et al. (2007)
	Conde Guadalupe	37.00	-4.80	350	868	1965–1982	Renschler et al. (1999)
	La Molina	42.40	2.00	1680	2172	1991–2000	Catari and Gallart (2010)
	La Tejería	42.75	-2.00	550	1000	1996–2005	Casali et al. (2008)
	Montblanc	41.38	1.16	400	1785	1992–2007	Gazquez et al. (2002)
	Real de San Vicente	40.14	-4.69	454	1450	1951–1970	ICONA (1988)
	Soutelo	42.50	-8.40	800	3020	1951–1970	Fernandez et al. (2010)
	Valencia	39.47	-0.37	15	1750	1983–2002	Roldán Soriano (2009)
	Verin	42.00	-7.90	373	1000	1951–1970	Fernandez et al. (2010)
Switzerland	Frienisberg	47.00	7.40	400	1200	1961–1990	Arranged from Ledermann et al. (2010)
Tunisia	Bargou	36.00	9.50	600	680	1970–1994	Hamed et al. (2002)
Turkey	Ankara	39.95	32.85	100	692	1961–1990	Doğan (1987)
	Istanbul	41.00	29.10	799	740	1961–1990	Özhan et al. (2005)
	Develi	38.40	35.50	1173	632	1980–1999	Irvem et al. (2007)
	Goksun	38.50	36.50	1180	1338	1980–1999	Irvem et al. (2007)
United Kingdom	Bala	53.00	-3.50	150	1200	1990–2000	Arranged from Davison et al. (2005) and Boardman (1993)

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