Spatiotemporal variation and trends in rainfall erosivity in China's dryland region during 1961–2012

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ABSTRACT

This study evaluated rainfall erosivity and its changes in China's dryland region during 1961–2012, using a daily rainfall erosivity model, ArcGIS based spatial interpolation tools, and the Mann–Kendall and Sen's methods. It was found that mean rainfall erosivity was $2\text{--}4098 \text{ MJ mm ha}^{-1} \text{ h}^{-1} \text{ a}^{-1}$ in the region, with the regional average of $794 \text{ MJ mm ha}^{-1} \text{ h}^{-1} \text{ a}^{-1}$, showing a significant increase trend accordingly with an increase in rainfall from northwest to southeast. Rainfall erosivity varied annually with the CV of 32.4–491.3%, and varied seasonally with about three-quarter distributed in summer. During 1961–2012, a statistically significant increase/decrease trend in rainfall erosivity was detected only for 24/10 stations or about 4.6/0.4% of the region, and a non-significant increase/decrease trend was observed for 32/45 stations or 16.5/10.0% of the region at the 50 to 90% confidence level. No statistically confidential trend was found for other 126 stations or 68.5% of the region. At regional level, mean annual erosivity of the arid zone indicated an upward trend, while the sub-humid zone a downward, but the semi-arid zone generally had no evidential trend. Comparing the 26 year's averages during 1987–2012 and 1961–1986, mean erosivity of the arid zone increased by 17.4%, the sub-humid zone decreased by 6.1%, and the semi-arid zone reduced slightly by 0.45%. The results suggest that change in the rainfall characteristics during the past half century had no significant influence on water erosion in the majority of the region, but it could probably induce an increase of water erosion risk in parts of the drier west and a slight decrease in parts of the wetter east including the Loess Plateau.

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1. Introduction

Water erosion causes environmental degradation (Lee and Heo, 2011; Xin et al., 2011) and is affected by natural and human factors involved in climate, terrain, soil, land use and management. During the past decades, the global and regional climate has shown various changes (IPCC, 2013). For precipitation, various changes have been observed in the amount, frequency and intensity (Back et al., 2011; Petkovšek and Mikoš, 2004). These changes influence erosivity of rainfall, and thus water erosion (Bazzano et al., 2010; Bonilla and Vidal, 2011; Busnelli et al., 2006). In the context of climate change, the effect of altered rainfall characteristics on soil erosion is one of the main concerns of soil conservation studies (Angulo-Martinez and Begueria, 2009).

Various indices have been developed to quantify rainfall erosivity, among which the R-factor of the Universal Soil Loss Equation (USLE) or its revised version, RUSLE (Renard et al., 1997), is the most widely used one (Diodato and Bellocchi, 2009). The R-factor is calculated

from the EI_{30} index using continuous rainfall data series with a time resolution of at least 15 min, and needs more than 20 years of rainfall data to obtain the mean annual value as to lessen the impact of natural climatic variations (Weiss, 1964; Wischmeier and Smith, 1958, 1978). Due to lack of rainfall process data, this method is seldom used to estimate rainfall erosivity in long-term trend or at large scale (Meusburger et al., 2012). As an alternative, simple algorithms have been used to predict rainfall erosivity from routine meteorological records of annual (Lee and Heo, 2011), monthly (Loureiro and Coutinho, 2001; Renard and Freimund, 1994), and daily (Yu and Rosewell, 1996; Zhang et al., 2002) rainfall data. In these algorithms, rainfall erosivity is calculated as an exponential (Richardson, 1983), linear (Lo et al., 1985) or logarithmic (Wischmeier and Smith, 1965) function of rainfall amount. Daily rainfall data can adequately present the rainfall variation characteristics and are usually available, and thus are preferably used to estimate rainfall erosivity (Angulo-Martinez and Begueria, 2009).

Detection of long-term trend in rainfall erosivity can provide information on the potential influence of rainfall changes on soil erosion. It is particularly useful for dryland areas, which are more sensitive to water erosion, and to climate changes (Feng and Fu, 2013; Huang et al., 2012) because of their fragile biophysical conditions. In China's dryland region, various changes have been observed in temperature,

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precipitation and other climatic factors during the past decades (Fan et al., 2012; Ge et al., 2013; Meng et al., 2013; Wang et al., 2006; Zhao, 2013). Many studies analyzed the effect of recent climate changes on droughts (Yao et al., 2013), stream runoff (Wang et al., 2012; Zhang et al., 2013), plant phenological development (Dai et al., 2012, 2014; Ge et al., 2013), and plant productivity (He et al., 2014; Xiao et al., 2013; Yao et al., 2012) in parts of the region. Some studies analyzed changes in rainfall erosivity, indicating that the rainfall erosivity tended to increase in north Xinjiang and north Tibet Plateau, but to decrease in the Loess Plateau (Liu et al., 2013; Xin et al., 2011; Zhang et al., 2003). These studies provided useful information on the variation of rainfall erosivity, but a further analysis is necessary for the region as a whole. This study has three aims: the first is to estimate the rainfall erosivity from daily rainfall data, the second is to analyze its spatiotemporal variation, and the third is to detect the trend of annual erosivity during 1961–2012. Spatial interpolation and GIS methods were used to map the spatial distribution of rainfall erosivity in China's dryland region.

2. Data and methods

2.1. Study area

The dryland region of China was delineated approximately based on the 600 mm isohyet of annual precipitation, taking into account the integrity of terrain units such as the Loess Plateau, a well-known region by its severe water erosion. The region extends between 69.4–128.0°E and

27.2–51.1°N with an area of 5.58 million km², about 60% of China's territory. It covers the Tibet Plateau, the Loess Plateau and the Inner Mongolia Plateau, involved in 12 provincial administrative units including the whole/most part of Xinjiang, Qinghai, Xizang (Tibet), Ningxia, Inner Mongolia, Gansu, Shaanxi, Shanxi, Hebei and Beijing, and a small part of Henan, Jilin and Liaoning (Fig. 1).

The region has a monsoon climate characterized by a distinct seasonal variation. The mean annual temperature is between −3.7 and 13.6 °C, and the regional average is 7.1 °C with the monthly temperature ranged from 21.0 °C in July to −9.4 °C in January. The mean annual precipitation is 33–623 mm, of which 61.8–76.1% is distributed in the rainy season from June to September. Based on the aridity index (K), the ratio of annual precipitation to annual potential evapotranspiration (Ci and Wu, 1997; Thornthwaite, 1948), the study area was divided into arid (K < 0.2), semi-arid (0.2 ≤ K < 0.5) and sub-humid (K ≥ 0.5) zones (Fig. 1). The arid zone including the hyper-arid has an annual precipitation generally less than 200 mm, covering the major part of Xinjiang, northwestern Tibet Plateau and western Inner Mongolia. The semi-arid zone has an annual precipitation normally between 200 and 450 mm, covering south central Tibet Plateau, northwestern Loess Plateau, eastern Inner Mongolia, and northern Xinjiang. The sub-humid region is distributed in the eastern, northeastern and southeastern part with an annual precipitation mostly above 450 mm, including east and south parts of the Loess Plateau, southeastern Tibet Plateau, and northern part of the North China Plain. Water erosion is serious in the hilly areas of the sub-humid and semi-arid zones, but it is weak in the

Fig. 1. Location of meteorological stations, major land units and climate zones in the study area.

arid zone particularly the hyper-arid part including the Taklimakan Desert, where wind erosion and soil salinization are the major problems of land degradation.

2.2. Rainfall data and the homogeneity check

The collected data include: 1) daily precipitation data during 1961–2012 from the National Meteorological Information Centre of China (<http://cdc.cma.gov.cn>) for 298 stations in the region (Fig. 1); and 2) 60-min rainfall data during 1981–1988 at 13 representative stations (Table 1; Fig. 1), and 30- and 60-min rainfall data for 599 storms at 79 sites during 1981–1988 from the China's Hydrological Yearbook, which are used to evaluate the estimated rainfall erosivity from daily data.

In this study, the evaluation of rainfall erosivity was based only on rainfall data, and thus the snow and hail records were excluded from the daily observation data at all stations. The years with one or more months of missing records were excluded for the analysis. By a systematic check using a batch program based on the Matlab software, 237 stations had complete records for the study period from 1961 to 2012, and 8 for 50–51 years, 12 for 40–48 years, 12 for 30–39 years, 24 for 20–28 years, 4 for 10–14 years and 1 for 7 years. The stations with observation span of less than 20 years are located in the arid western part where meteorological stations are sparsely distributed.

To detect trend in rainfall erosivity, the rainfall data should be consistently recorded at a station without changes in the location and the general surroundings, the observation equipment, and the method of data collection during the observation period, as the changes can result in inhomogeneity of data records (Khaliq and Ouada, 2007). In this study, the Standard Normal Homogeneity Test (Alexandersson, 1986), a location specific homogeneity test, was applied to examine the consistency of rainfall data. This test assumes that the data collected at all sites within a same climatic region should be highly correlated, have similar variability, and differ only by scaling factors and random sampling variability (Costa and Soares, 2009), therefore, homogeneity of a data series can be checked by comparing with a single or multiple reference series (Toreti et al., 2011). For rainfall records at each of the 298 stations, the homogeneity was examined by comparing the annual rainfall value with four highly correlated reference series from the surrounding stations, following the procedure in literature (Jiang et al., 2008; Khaliq and Ouada, 2007; Liu and Sun, 1995). In total, 290 stations have passed the homogeneity test, of which 237 stations have a similar duration of 50–52 years, and were used for the trend analysis of annual erosivity.

2.3. Model of rainfall erosivity

Annual rainfall erosivity was estimated for each of the 298 stations from daily rainfall data using the method of Zhang et al. (2002), which was well tested and widely used in China (Xin et al., 2011; Zhang et al., 2003). This method was derived from the Richardson (1983)

Fig. 2. Relationship between coefficients α and β .
Data source: Zhang et al., 2002.

equation, $EI = \alpha P^\beta$, where EI is an event's erosivity and P is the rainfall, with the equation below to calculate half-month rainfall erosivity, R ($\text{MJ mm ha}^{-1} \text{ h}^{-1}$):

$$R = \alpha \sum_{j=1}^k (P_j)^\beta \quad (1)$$

where P_j is the erosive rainfall (mm) for day j in a given half-month of k days, and α and β are coefficients. For daily rainfall equal to or larger than the threshold value of 12 mm, P_j is the actual rainfall, otherwise P_j equals 0 (Xie et al., 2000). The annual erosivity is the summation of all half-month erosivity during the year.

The β varied with changes in rainfall characteristics, and was estimated with the empirical equation below (Zhang et al., 2002):

$$\beta = 0.8363 + \frac{18.144}{P_{d12}} + \frac{24.455}{P_{y12}} \quad (2)$$

where P_{d12} and P_{y12} are the average daily and annual rainfall for days with rainfall ≥ 12 mm, respectively.

Zhang et al. (2002) formulated an exponential equation to obtain the α value from the estimated β , based on the data obtained from 71 representative sites in whole China. In this study, we selected the α and β data pairs at 33 stations that are located in our study area (Fig. 2), and then reformulated the equation in the form below:

$$\alpha = 17.554 \beta^{-7.042} \quad (3)$$

Table 1

General information of 13 stations at which 60-min rainfall data were collected for 1981–1988.

Station	Lon. (°)	Lat. (°)	Elevation (m)	Annual precipitation (mm)
Yamadu	81.80	43.60	770	237
Xinquman	82.72	41.03	1029	60
Bayinbuluge	84.13	43.00	2435	253
Dangchengwan	94.90	39.50	1375	160
Xiangride	98.13	35.75	3441	260
Dongjiazhuang	101.27	36.67	2636	419
Shiyanghe	102.73	38.18	1470	136
Renda	105.75	35.17	1380	462
Qingyang	107.88	36.00	1096	489
Huangling	109.27	35.58	829	594
Longtougua	109.78	40.30	1050	260
Qingshuihe	111.68	39.90	1096	365
Chakou	111.78	37.63	1120	490

Table 2

General description of validation statistics.

- 1) Mean relative bias error: $MRBE = \frac{\sum_{i=1}^n (D_i - C_i)}{\sum_{i=1}^n C_i} \times 100$
- 2) Mean relative absolute error: $MRAE = \frac{\sum_{i=1}^n |D_i - C_i|}{\sum_{i=1}^n C_i} \times 100$
- 3) Root mean square error: $RMSE = \sqrt{\frac{\sum_{i=1}^n (D_i - C_i)^2}{n}}$
- 4) Coefficient of determination: $R^2 = \left[\frac{\sum_{i=1}^n (D_i - MD) \times (C_i - MC)}{\sqrt{\sum_{i=1}^n (D_i - MD)^2 \times \sum_{i=1}^n (C_i - MC)^2}} \right]^2$

Where n is the number of years, D_i is the erosivity in year i and MD the average annual erosivity from daily rainfall data, and C_i is the erosivity in year i and MC the average annual erosivity from 60-min rainfall data.

Fig. 3. Relationship between event's erosivity values of EI and $(EI)_{60}$ (a) and the adjusted $(EI)_{60}$ (b) for the 599 storms.

2.4. Model evaluation

The daily erosivity model was evaluated at 13 representative stations (Table 1), with four validation statistics (Table 2), including mean relative bias error (MRBE), mean relative absolute error (MRAE), root mean square error (RMSE), and coefficient of determination (R^2).

The yearly erosivity for the model validation was calculated from the classic EI index approach using 60-min interval data of erosive rainfall events. With this coarse time resolution data, the rainfall intensity and thus the storm erosivity are often substantially underestimated. Therefore, many researchers have suggested a conversion coefficient to reduce the bias error (Renard et al., 1997; Weiss, 1964; Zhu et al., 2011) based on higher time resolution data from continuous records of rainfall intensity. Denoted this coefficient as c , the individual storm erosivity, EI, was computed with the equation below:

$$EI = c(EI)_{60} \quad (4)$$

where $(EI)_{60}$ is the erosivity of an erosive event with the accumulative rainfall of larger than 12.7 mm (Renard et al., 1997), as computed from 60-min interval data. The variable I (mm h^{-1}) is the maximum 30-min rainfall intensity, taking the maximum hourly rainfall of the storm, and E (MJ ha^{-1}) is the total kinetic energy, as calculated using the Brown and Foster equation (1987) in the form below (Forster, 2004):

$$E = \sum_{k=1}^m 0.29[1 - 0.72 \exp(-0.082i_k)](\Delta V)_k \quad (5)$$

where m is the number of 60-min intervals during an erosive rainfall event, i_k is the rainfall intensity (mm h^{-1}), and ΔV_k is the rainfall depth (mm) during time period k . The annual erosivity is the summation of all event's erosivity occurred during the year. The coefficient c was obtained by linear regression, based on the event's erosivity values of 599 storms, as calculated from 30- and 60-min interval data, respectively.

2.5. Mean annual rainfall erosivity and its spatiotemporal variation

Mean annual rainfall erosivity at each of the 298 sites was simply calculated by averaging the annual values of observation years, without considering difference in the duration. Based on these site's values, the spatial distribution of mean annual erosivity was interpolated with the high accuracy surface modeling (HASM) tool (Yue et al., 2004). Different from the Kriging and Inverse Distance Weighted (IDW) methods with an assumed homogeneous stochastic field (Yue, 2010), the HASM takes into account the non-homogeneity caused by topographic and locational factors, and thus has a relatively low reliance on the information of sampling points (Zhao, 2013). In this study, the rainfall erosivity at all stations was interpolated by a quadratic polynomial equation, taking into account the influence of latitude, longitude, elevation and slope. A stepwise regression was used to select the independent variables, and a multicollinearity analysis of the variables was conducted to ensure that there is no linear relationship between two selected variables. The residual was rectified by the HASM model with fitting result and information of the sites. Based on the interpolated

Table 3

Annual rainfall erosivity ($\text{MJ mm ha}^{-1} \text{h}^{-1} \text{a}^{-1}$) from daily and 60-min rainfall data and the validation results.

Hydrological site	Erosivity from daily data		Erosivity from 60-min data		R^2	MRBE (%)	MRAE (%)	RMSE ($\text{MJ mm ha}^{-1} \text{h}^{-1}$)
	Mean	Range	Mean	Range				
Yamadu	65	0–159	77	0–191	0.93	−15.84	21.3	24.8
Xinquman	59	21–176	66	0–356	0.82	−10.30	35.1	32.1
Bayinbuluge	91	0–173	87	7–445	0.88	4.48	23.8	23.1
Dangchengwan	232	232–472	203	16–252	0.81	14.63	38.6	90.7
Xiangride	70	0–118	62	7–221	0.46	12.10	34.6	32.0
Dongjiazhuang	407	111–604	297	82–483	0.88	30.67	22.4	107.4
Shiyanghe	77	77–110	60	24–56	0.94	21.67	36.9	22.7
Renda	707	167–1249	654	48–1111	0.96	11.41	13.6	113.3
Qingyang	1018	558–2413	1002	342–2340	0.92	4.78	10.0	125.6
Huangling	1465	570–2456	1659	759–3612	0.86	2.77	14.9	230.0
Longtougua	532	83–980	624	188–1175	0.93	−12.31	15.5	126.4
Qingshuihe	627	232–1008	601	194–1061	0.97	7.54	11.3	96.0
Chakou	1279	723–2239	1021	647–1729	0.84	28.60	22.8	347.2

Fig. 4. Spatial distribution of mean annual rainfall erosivity in the study area.

results, the regional average of mean annual erosivity was obtained by averaging the grid value using the Geostatistical Analyst tool in ArcGIS.

For a specific year, the regional mean value was calculated by averaging the yearly erosivity at the 237 stations that have same time span from 1961 to 2012. Seasonal (monthly) distribution of annual erosivity was expressed in percent of seasonal (monthly) erosivity as the annual total. Annual variation was indicated by coefficient of variation (CV), presented as a percentage of the standard deviation of annual erosivity to the mean value of observation years. The spatial distribution of CV was interpolated with the IDW approach based on the values at the 290 stations with homogeneous data records. Compared to the interpolation result by Kriging method, the IDW described the spatial variation comparatively well in this study.

2.6. Trend analysis and estimation of change rate

The trend of annual rainfall erosivity was detected with the nonparametric Mann–Kendall test, and the trend magnitude was estimated

with the nonparametric Sen's method (Gilbert, 1987). The evaluation was based on the Excel template application MAKESENS, as developed by Salmi et al. (2002) for estimating trends in the time series of annual atmospheric and precipitation pollutant concentrations.

The Mann–Kendall test is widely used to identify trends in time series (Kendall, 1975; Mann, 1945), such as long-term climatic and hydrological data (Beran et al., 1996; Burn and Elnur, 2002; Sicard et al., 2010). It is based on two statistic variables S and Z to determine whether a time series of n data points has a significant trend. The S is the summation of $\text{sgn}(x_{i+1} - x_i)$ that has a value of 1 for data point value x_i is less than x_{i+1} , 0 for $x_i = x_{i+1}$, and -1 for $x_i > x_{i+1}$ ($i = 1, 2, \dots, n-1$). The Z is the standardized test statistic of S , which equals zero for $S = 0$, $(S - 1) / [\text{var}(S)]^{0.5}$ for $S > 0$, and $(S + 1) / [\text{var}(S)]^{0.5}$ for $S < 0$, where $\text{var}(S)$ is the variance of S (Salmi et al., 2002). A positive (negative) value of Z indicates an upward (downward) trend. At α level of significance, a null hypothesis is rejected if the absolute Z value is greater than $Z_{1-\alpha/2}$, where $Z_{1-\alpha/2}$ is obtained from the standard normal cumulative distribution tables. The Sen's method is a consistent nonparametric estimator for the coefficient of a linear regression (trend magnitude),

Table 4

Regional mean annual rainfall and erosivity for the whole region and the three climatic zones.

Zones	Rainfall (mm)	Erosive rainfall (mm)	Rainfall erosivity (MJ mm ha ⁻¹ h ⁻¹)	Ratio of erosivity to rainfall
Arid zone	80	24	97	1.21
Semi-arid zone	283	114	407	1.44
Sub-humid zone	483	267	1372	2.84
Whole region	333	172	794	2.38

Table 5

Seasonal distribution (%) of regional mean annual erosivity for the whole region and the three zones.

Climate zone	Spring	Summer	Autumn
Arid zone	10.5	78.2	11.1
Semi-arid zone	9.7	76.1	13.0
Sub-humid zone	8.4	76.0	14.9
The whole region	9.2	76.7	13.6

i.e., the slope measured as a change in unit time (Sen, 1968). The change rate in this study was expressed in a percentage of the slope value as the mean erosivity of observation years.

The Z value and the change rate of annual erosivity were estimated for each of the 237 stations with homogenous rainfall records using the MAKESENS. The estimated Z values at each site were imported into ArcGIS, and then interpolated using the IDW method to reveal the spatial variation of trends.

3. Results

3.1. Model validation results

To determine the coefficient c in Eq. (4), two rainfall erosivity values, EI and $(EI)_{60}$, were calculated for each of the selected 599 storms from 30- and 60-min interval rainfall data, respectively. The results were presented in Fig. 3a, indicating that the $(EI)_{60}$ value was 8–2094 MJ mm ha⁻¹ h⁻¹, systematically lower than the EI value of 19–3234 MJ mm ha⁻¹ h⁻¹. Compared to the mean EI value of all storms (351 MJ mm ha⁻¹ h⁻¹), the mean $(EI)_{60}$ value (206 MJ mm ha⁻¹ h⁻¹) was underestimated by 43%. This systematic error was mainly caused by the bias error in the maximum 30-min rainfall intensity I , as the value of I for all storms from 60-min data was underrated, averagely by 39%, while the total energy E was slightly underestimated by only 3%. By linear regression, we obtained the regression slope, i.e., the coefficient c as 1.793 ($R^2 = 0.95$). This slope value is close to the results obtained by Yin et al. (2007) with breakpoint data taken from continuous rain gauge charts, e.g., the value was 1.811 at Zizhou station in the semi-arid northern Shaanxi of our study area. Adjusted with this c value, the storm's erosivity from 60-min rainfall data was greatly improved (Fig. 3b), with the absolute error (MRAE) decreased to 10.7% from 32.6%, and the root mean square error (RMSE) decreased to 7.0 MJ mm ha⁻¹ h⁻¹ from 18.1 MJ mm ha⁻¹ h⁻¹.

Taking 1.793 for the coefficient c , the annual erosivity during 1981–1988 was computed using Eq. (4) from 60-min rainfall data for the 13 representative stations, used to evaluate the estimated results with Eq. (1) from daily rainfall. The validation results were shown in Table 3. For the mean annual erosivity of 8 years, a linear regression indicated that both values at the 13 sites were highly correlated ($R^2 = 0.96$) with the slope of 0.997, close to 1. The mean bias error (MRBE) was between -15.8 and 14.6% at 10 sites, and 21.7–30.7% at 3 sites. For the yearly erosivity, the predicted value from daily rainfall data was well related to that from 60-min data ($R^2 = 0.82$ –0.97) at the stations except Xiangride, but the MRAE and RMSE were rather big (Table 3). In general, the erosivity predicted from the daily model was plausible for most sites.

3.2. Spatial variation in annual rainfall erosivity

The mean annual erosivity at the 298 stations was between 2 and 4098 MJ mm ha⁻¹ h⁻¹, and was less than 500 MJ mm ha⁻¹ h⁻¹ in 65% of the region, showing a zonal variation characteristic and an increasing trend from northwest to southeast (Fig. 4). The erosivity was significantly smaller, normally below 100 MJ mm ha⁻¹ h⁻¹ a⁻¹ in the arid northwestern part, while evidently larger, generally above 1000 MJ mm ha⁻¹ h⁻¹ a⁻¹ in the sub-humid eastern part including the most of the Loess Plateau. In other parts including the Tibet Plateau and the north central Inner Mongolia Plateau, the erosivity was mostly between 100 and 500 MJ mm ha⁻¹ h⁻¹ a⁻¹. For the whole region, the regional average was 794 MJ mm ha⁻¹ h⁻¹ a⁻¹, and for the arid, semi-arid and sub-humid zone, it was 97, 407 and 1372 MJ mm ha⁻¹ h⁻¹ a⁻¹, respectively. Based on the term of erosivity density (Forster, 2004), we calculated the ratio of erosivity to rainfall and found that it increased from the arid to sub-humid zone with an increase of annual rainfall (Table 4).

3.3. Seasonal and annual variation in rainfall erosivity

The rainfall erosivity showed a great seasonal and annual variation. Influenced by the monsoon climate, the erosivity was highly concentrated in summer season from June to August. For the whole region, the mean summer erosivity occupied 76.7% of the annual total, and for the arid, semi-arid and sub-humid zones, it covered 78.2, 76.1 and 76.0%, respectively, showing a rather similar seasonal distribution in the three zones (Table 5). As rainstorms did not fall in winter and seldom occurred in spring, the erosivity showed a great monthly variation, with the maximum erosivity generally occurred in July or August in the semi-arid and sub-humid zones, and in June or July in the arid zone (Fig. 5). A trend analysis with the Mann–Kendall revealed that the proportion of summer erosivity for the whole region tended to decrease from 1961 to 2012 ($Z = -1.55$).

Annual variation of rainfall erosivity (CV) at site level varied from 32.4 to 491.3% across the region, generally decreased from northwest to southeast (Fig. 6). In the arid zone where erosive rainfall events occurred only in few years, the CV was markedly greater, mostly exceeding 110%. In the semi-arid zone, the CV ranged 40–160%, generally between 50 and 100% at most sites. In the sub-humid zone, the CV was much lower, mostly less than 60%. For the entire region and the three zones, the mean annual erosivity also showed an apparent variation among years, even among the three-years moving averages (Fig. 7). The mean erosivity in the entire region was 511 to 1200 MJ mm ha⁻¹ h⁻¹ a⁻¹ during 1961–2012, and the CV was 19.2%, close to that of 20.1% in the semi-

Fig. 5. Distribution (%) of monthly rainfall erosivity at representative stations in arid zone (a), semi-arid zone (b) and sub-humid zone (c). The annual precipitation at the selected stations (for locations see Fig. 4) was 33–41 mm in the arid zone, 192–432 mm in the semi-arid zone, and 495–562 mm in the sub-humid zone.

Fig. 6. Spatial distribution of relative variability (CV) of annual erosion in the study area.

arid zone and 20.6% in the sub-humid zone, but much lower than that of 35.7% in the arid zone.

3.4. Trend and change rate in annual rainfall erosivity

The Z test value of annual rainfall erosivity during 1961–2012 was between -2.47 and 4.06 for the 237 stations. A total of 134 stations showed a positive Z value and 103 a negative value. A spatial statistics with ArcGIS indicated that the stations in the arid zone mostly had a positive Z value, while in the sub-humid zone a negative value, demonstrating a distinct regional difference in the direction of possible trends. In the semi-arid zone, the number of stations with a positive and negative Z value did not show an apparent difference. A further analysis (Table 6) indicated that the absolute Z value at 126 stations was below the value of 0.675 at the 0.5 significance level, implying that the possibility of a trend presence was below 50%. At 24 (10) stations, the absolute Z value exceeded the threshold value of 1.645 at the 0.1 significance level, indicating a significant increase (decrease) trend in annual erosivity. At 32 (45) stations, the absolute Z value was 0.675–1.645, showing a non-significant increase (decrease) trend at the 0.5 to 0.1 significance level. At the 24 (10) stations, the change rate of annual erosivity was 2.2 to 10.1% (-2.6 to -7.3%) per decade, and at the 32 (45) stations, it was 6.6 to 26.3% (-5.5 to -9.5%) per decade.

Fig. 8 indicated that in the majority (68.5%) of China's dryland region, no trend was detected for the annual erosivity at the above 50% confidence level, as the absolute Z value was less than 0.675. In 21.1% of the region, the annual erosivity showed a non-significant increase trend ($Z = 0.675$ – 1.645), and in 10.0% a non-significant decrease

trend ($Z = -0.675$ to -1.645). A statistically significant upward and downward trend was observed only in a limited area, occupying 4.6% and 0.4% of the region, respectively (Table 6). The trend occurrence showed a highly concentrated distribution characteristic: the area with upward trends was mainly distributed in the middle part (semi-arid to arid transition zone), the northern Xinjiang, and central Tibet, while the area with downward trends mainly distributed in the sub-humid zone including southern Loess Plateau (Fig. 8). Few sites showed a different trend from the surroundings, probably due to the influence of high mountains.

In the arid zone, the regional mean annual erosivity was detected a significant increase trend at the 0.01 significance level ($Z = 2.83$), with an increasing rate of $7.1 \text{ MJ mm ha}^{-1} \text{ h}^{-1}$ or 8.3% per decade. In the sub-humid zone, it exhibited a non-significant decline trend ($Z = -1.43$) at a decrease rate of $41.4 \text{ MJ mm ha}^{-1} \text{ h}^{-1}$ or 3.3% per decade. No trend at an acceptable confidence level was found for the mean annual erosivity of the semi-arid zone ($Z = -0.29$) and the whole region ($Z = -0.41$). A further analysis by comparing the mean erosivity values during the first 26 years of 1961–1986 and the second of 1987–2012 indicated a similar trend. During 1987–2012, the mean annual erosivity of the arid zone ($100 \text{ MJ mm ha}^{-1} \text{ h}^{-1}$) markedly increased by 17.4%, the sub-humid zone ($1389 \text{ MJ mm ha}^{-1} \text{ h}^{-1}$) decreased by 6.1%, and the semi-arid zone ($702 \text{ MJ mm ha}^{-1} \text{ h}^{-1}$) slightly decreased by 0.45%, compared to that during 1961–1986. For the whole region, the mean erosivity during 1987–2012 was $775 \text{ MJ mm ha}^{-1} \text{ h}^{-1} \text{ a}^{-1}$, 3.6% lower than the value of $805 \text{ MJ mm ha}^{-1} \text{ h}^{-1} \text{ a}^{-1}$ during 1961–1986. Similar change in erosivity was also found for the Loess Plateau by Elbasit et al. (2013).

Fig. 7. Annual variation of rainfall erosivity for the whole region (a), arid zone (b), semi-arid zone (c) and sub-humid zone (d).

4. Discussion

This study presented the spatial, annual and seasonal variation of rainfall erosivity in China's dryland region based on the modified Richardson (1983) approach by Zhang et al. (2002). The results showed that the rainfall erosivity varied greatly over spatial, increased from the mean value below 100 MJ mm ha⁻¹ h⁻¹ a⁻¹ in the arid zone to 1000–4098 MJ mm ha⁻¹ h⁻¹ a⁻¹ in the sub-humid zone. This spatial distribution was in accordance with the results of Liu et al. (2013), which indicated that the annual erosivity was generally lower than 100 MJ mm ha⁻¹ h⁻¹ in the arid region and 1000–4000 MJ mm ha⁻¹ h⁻¹ in the sub-humid region of China. In the Loess Plateau, the annual rainfall erosivity ranged from 500 to 2000 MJ mm ha⁻¹ h⁻¹, with the regional average of 1018 MJ mm ha⁻¹ h⁻¹, rather close to the average of 1276 MJ mm ha⁻¹ h⁻¹ during 1956–2008, as obtained by Xin et al. (2011). In the Tibet Plateau, the mean annual erosivity was generally lower than 500 MJ mm ha⁻¹ h⁻¹, similar to the results of Zhang et al. (2003). Annual rainfall erosivity varied from 679 to 2428 MJ mm ha⁻¹ h⁻¹ in Hebei province, showing a rather big difference from the values between 930 and 1667 MJ mm ha⁻¹ h⁻¹

calculated by the revised Fournier index during 1961–2000 (Men et al., 2008). The spatial variation of rainfall erosivity was largely determined by the spatial distribution of annual rainfall. Regression results using the average values at the 298 stations indicated that the mean annual erosivity was significantly related to the mean annual erosive rainfall ($R^2 = 0.91$) and annual total rainfall ($R^2 = 0.58$).

The erosivity showed a distinct seasonal and monthly variation because rainfall particularly the erosive rainfall mostly falls in summer season, explaining why water erosion occurs mainly in summer months in the region. With an increase in rainfall, the annual variation of rainfall erosivity decreased from the drier west to the wetter east, showing a negative relationship with annual rainfall. This was because higher rainfall is generally associated with a lower annual variation, and thus a lower variation in annual erosivity, as annual erosivity was highly related to annual rainfall. Based on the regional yearly values during 1961–2012, the regression analyses indicated that the annual rainfall erosivity was significantly related to the annual erosive rainfall ($R^2 = 0.93$), the annual total rainfall ($R^2 = 0.82$) and the mean rainfall intensity, the ratio of annual rainfall to rain days ($R^2 = 0.70$). Similar relationships were found at site level (Table 7). In general, the annual variation

Table 6

The Z test values and trend of annual rainfall erosivity during 1961–2012 for the 237 stations.

Class	Z value	Trend	Number of stations				% as the study area
			Arid zone	Semi-arid zone	Sub-humid zone	Total	
1	≥1.645	Significant increase	5	17	2	24	4.6
2	0.675–1.645	Non-significant increase	7	14	11	32	16.5
3	–0.675–0.675	No trend	39	32	55	126	68.5
4	–1.645 to –0.675	Non-significant decrease	1	11	33	45	10.0
5	≤–1.645	Significant decrease	0	3	7	10	0.4
Total	–2.470–4.061		52	77	108	237	100.0

Fig. 8. Spatial variation of changing trend in the annual rainfall erosivity.

of rainfall erosivity was much bigger than that of annual rainfall, as not all rainfall events have erosive power and the erosivity is often generated by few heavy storms, particularly in the arid areas.

Most stations in the drier west part including the most of Tibet Plateau showed a positive Z value, while a negative value in the wetter east part including southern Loess Plateau, indicating an apparent spatial pattern in the trend occurrence of erosivity. However, a statistically significant trend in erosivity was detected only for 34 stations or about 5% of the region, and a non-significant trend was observed for 77 stations at the 50 to 90% confidence level. No statistically confidential trend can be determined for other 126 stations or in 68.5% of the region. At regional level, the mean annual erosivity was observed a significant increase trend for the arid zone, and non-significant decrease trend for the sub-humid zone, while no evidential trend for the semi-arid zone. For the whole region, the mean erosivity showed a non-significantly decreasing trend during 1961–2012 ($Z = -1.38$), in accordance with the trend of annual erosive rainfall ($Z = -0.69$) and annual total rainfall ($Z = -0.56$), but different from the trend of mean rainfall intensity ($Z = 0.81$) that showed a non-significantly increasing trend. Similar trends in annual rainfall erosivity during the past 50 years were found

by Liu et al. (2012, 2013), i.e., an increase trend in the arid north central Tibet Plateau and northern Xijiang, and a decrease trend in the sub-humid Bohai area. For the Loess Plateau as a whole, the mean erosivity was also found to decrease during 1956–2008 (Xin et al., 2011), and reduced by 10% between 1960s and 1990s (Abd Elbasit et al., 2013).

The results of this study suggest that the change in rainfall characteristics during the past 52 years could have no significant influence on water erosion risk in the most parts of the region. However, it probably had a negative impact for the arid areas, as the regional mean erosivity during 1987–2012 was evidently increased by 17.4% compared to that during 1961–1986, while a positive effect for the sub-humid part including the Loess Plateau, since it decreased by 6.1%. These changes in rainfall erosivity could probably induce a corresponding slight change in the water erosion risk.

Uncertainties could be involved because the model may have rather big bias errors in the predicated erosivity for some stations particularly in the arid areas. The relief and elevation differ greatly across the region, and the impact on erosivity was not sufficiently considered. Another factor is that the stations with available data are very limited and sparsely distributed in the arid area and high mountainous areas (e.g., the Tibet Plateau), which could cause errors for the interpolated results.

5. Conclusions

Based on a simple daily erosivity model, this study provided insights into the spatial, inter-annual and seasonal variation of rainfall erosivity during 1961–2012 in China's dryland region. The Mann–Kendall and Sen's methods were used to detect the change trend in rainfall erosivity

Table 7

The coefficient of determination (R^2) of annual rainfall erosivity related to annual total and erosive rainfall, and rainfall intensity during 1961–2012 at 5 representative stations.

R^2	Annual rainfall	Erosive rainfall	Rainfall intensity
Beijing	0.87	0.90	0.76
Taiyuan	0.75	0.88	0.70
Lanzhou	0.55	0.70	0.57
Hohhot	0.78	0.84	0.73
Linhe	0.48	0.71	0.58

at site and regional scale, to reveal the potential impact of climate change on water erosion risk. Main findings were summarized below:

The long-term average of annual erosivity varied greatly from 2 to 4098 MJ mm ha⁻¹ h⁻¹ a⁻¹, with the annual variation (CV) ranging from 32.4 to 491.3% across China's dryland region. The spatial distribution of erosivity was in accordance with the spatial change of annual rainfall particularly erosive rainfall. The mean annual erosivity for the whole region was 511–1200 MJ mm ha⁻¹ h⁻¹ during 1961–2012 with the average of 794 MJ mm ha⁻¹ h⁻¹. Rainfall erosivity had a distinct seasonal variation with about three-quarter of the annual total distributed in summer season.

During 1961–2012, total 34 stations were detected a significant trend in annual erosivity at the 0.1 significance level, comprising 10 stations with a downward trend and 24 stations with an upward trend. A total of 77 stations representing 26.5% of the study area demonstrated a non-significant trend in annual erosivity, and more than half stations with an area of 68.5% was not found a presence of trends at the confidence level of above 50%. In the drier west part, the annual erosivity mostly showed a positive Z value, while in the wetter east part, generally a negative value, indicating a distinct regional difference in the trend occurrence. The mean annual erosivity of the arid zone showed a significant increase trend, the sub-humid zone a non-significant decrease trend, and the semi-arid zone generally no trend. The mean erosivity during 1987–2012 increased by 17.4% for the arid zone, and decreased by 6.1% for the sub-humid zone, 0.45% for the semi-arid zone, and 3.6% for the whole region, compared to that during 1961–1986. These results suggest that change in rainfall characteristic during past half century had no significant influence on water erosion in the majority of the region, but could probably induce an increase in the risk of water erosion in parts of the drier west and a slight decrease in parts of the wetter east.

The study contributed to the understanding of spatiotemporal variation of annual rainfall erosivity and the possible impact of changes in the rainfall characteristics associated with global warming in China's dryland area during the past half century. Uncertainties could be involved because of limited data availability and bias errors of the model.

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