Catena 160 (2018) 103–111

Contents lists available at ScienceDirect

Catena

journal homepage: www.elsevier.com/locate/catena

Holocene erosion triggered by climate change in the central Loess Plateau of China

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A R T I C L E I N F O

Keywords:
Radiocarbon dating
Holocene climate
Historical erosion
Net primary productivity
Rainfall erosivity

A B S T R A C T

Understanding changes in Holocene erosion is essential for predicting soil erosion in the future. However, the quantitative response of natural erosion to Holocene climate change is limited for the Loess Plateau of China. In this study, two soil profiles were investigated on the Luochuan and Yanchang sites in the central Loess Plateau of China; and four climate indicators, i.e. magnetic susceptibility, calcium carbonate content, total organic carbon content, and clay content ($<$ 0.005 mm), were analyzed to describe climate change. The equations fitted using modern pedogenic susceptibility, precipitation, and temperature were used to quantitatively reconstruct paleoprecipitation and paleotemperature in the Holocene. The current relationship between soil erosion intensity and precipitation was determined and used to estimate historical erosion. Results indicated that climate was coldest and driest between 12,000 and 8500 cal. yr BP, and became warmer and wetter during 8500 to 5500 cal. yr BP. The warmest and wettest climate was from 5500 to 3000 cal. yr BP and was getting colder and dryer over the last 3000 years. Holocene erosion intensity changed with fluctuation of mean annual precipitation, and these changes were different on both sites. The peak erosion values were 20,966 $t \cdot km^{-2} \cdot yr^{-1}$ in 7500 cal. yr BP and 21,148 $t \cdot km^{-2} \cdot yr^{-1}$ in 3300 cal. yr BP on the Luochuan and Yanchang sites, respectively. Furthermore, more severe soil erosion with a faster increase was estimated on the Yanchang site than Luochuan site with a range between 6547 and 11,177 $t \cdot km^{-2} \cdot yr^{-1}$ during the last 1800 years. This study proposed a new method to quantify historical soil erosion triggered by climate change, which not only can derive detailed soil erosion intensity change with variation of climate, but also provide a way to compare soil losses between different areas.

1. Introduction

Soil is an important natural resource that humans rely on and civilization is based upon. The erosion of topsoil not only affects local agricultural and industrial productivity, but also causes serious offsite environmental problems, e.g. chemical pollution and river channel sedimentation. Soil erosion is usually determined by both natural conditions, e.g. rainfall, gradient, surface cover, and soil type, and anthropogenic activities, e.g. farming, grazing, and constructing. The Loess Plateau of China, located within the middle reaches of the Yellow River, is in the semiarid zone where natural conditions are highly susceptible to erosion, and human activities have increased during the Holocene (Ren and Zhu, 1994; Shi et al., 2002). Owing to a combination of natural and human-induced erosion, it became one of the most serious soil erosion areas in the world (Anonymous, 2004). Therefore, a primary objective of erosion control and soil conservation in the Loess Plateau should be to reduce total erosion to close to, or even lower than, the natural erosion rates. However, the rates of the natural erosion, which are mainly determined by the geological environment and climate (Zhang et al., 2001), are not constant and are difficult to predict (Mayewski et al., 2004). During the Holocene, considering the relative stability of the geological environment in the Loess Plateau of China, climate change played a dominant role in natural erosion (Shi et al., 2002; He et al., 2006). Therefore, for assessment and prediction of the natural erosion rates in this period, it is important to figure out its response to climate change.

The loess profile contains the most abundant information about the geologic evolution during the Quaternary period. It records the progression

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http://dx.doi.org/10.1016/j.catena.2017.09.013
Received 29 December 2016; Received in revised form 8 September 2017; Accepted 16 September 2017
Available online 02 October 2017
0341-8162/

of the paleoclimate, neotectonism, paleogeography, and other important geological events in the Quaternary period in the Mainland China. This profile also records the integrated processes during the evolution of global paleoclimate and paleoenvironment (Liu, 1985; An et al., 1990). Therefore, the loess plateau is one of the best geological information carriers for global change research because it provides precious and valuable conditions in spatial and temporal dimensions (Liu et al., 1986). Several researchers have conducted such studies in the Loess Plateau of China (Hovan et al., 1989; Chen et al., 1997; Porter et al., 2001; Maher et al., 2003; An et al., 2006; Stevens et al., 2006; Huang et al., 2009). They have investigated the evolution of paleoclimate and paleoenvironment to provide scientific basis for forecasting future climate evolution. Some quantitative methods were also developed to estimate historical soil erosion in the Loess Plateau of China. One method used was to calculate historical soil erosion intensity based on the speculation of gully volume (Bai, 1994; Sang and Gan, 2005). Another common method was to compute the soil loss from the Loess Plateau of China according to the sediment in the Yellow River delta, continental shelf of the Bohai Sea, and river terrace (Ren and Zhu, 1994; Shi et al., 2002; He et al., 2004). These methods were useful but could hardly provide information on the response of natural erosion to Holocene climate change.

Numerous studies (Kirkby and Cox, 1995; Istanbulluoglu and Bras, 2006; Benton and Newell, 2014) have shown that mean annual sediment yield is a function of mean annual precipitation in various areas. Although these functions varied, they showed similar changing patterns of the relationships between sediment yield and precipitation which were called the Langbein-Schumm curve (Langbein and Schumm, 1958). Those curves were primarily a function of climatic condition and land use (Collins and Bras, 2008). Although other factors, e.g. soil and topography, were also crucial in determining the absolute magnitude of sediment yield from drainage basins, they mainly affected the scatter of individual points around the curve. Without regard to the human activities, Xu (2005) found that the land use factor was mainly determined by natural vegetation which was expressed by the index of net primary productivity (*NPP*, $t \cdot ha^{-1} \cdot yr^{-1}$). In addition, the sediment delivery ratio in the Loess Plateau approaches 1 (Xu, 1999; Wei et al., 2006). Therefore, the Langbein-Schumm curve provides a possible way to evaluate past soil erosion rate if the mean annual precipitation and *NPP* are known.

The aims of this paper were to: (1) investigate the climate change in the central Loess Plateau of China during the Holocene; (2) reconstruct and assess paleoprecipitation and paleotemperature; (3) evaluate the response of soil erosion to Holocene climate change.

2. Material and methods

2.1. Study area and field work

The study area is located in the central Loess Plateau of China. The Loess Plateau in northwest China covers an area of 530,000 km^2; the loess deposits typically range in thickness from 30 to 80 m. Aeolian dust accumulation and soil erosion by overland flow, being the prevailing surface processes, have been forming the landscape of the Chinese Loess Plateau for the past 2.6 million years. The loessial soils are characterized by yellowish colors, absence of bedding, silty texture, looseness, and collapsibility when saturated. The Loess Plateau is conveniently divided into three zones: sandy loess in the northern part, silty loess in the middle, and clayey loess in the southern (Liu, 1985). The present-day vegetation in the study area is dominated by grasses and shrubs (*Poaceae*, *Leguminosae*, *Labiatae*, *Rhamnaceae*, and *Compositae*).

Two sites, Luochuan (N35°40′56.4″, E109°25′14.5″) and Yanchang (N36°38′36.3″, E109°55′12.8″) were selected for study (Fig. 1). The annual mean temperature is 9.2 °C and 8.9 °C and the annual rainfall is 622 mm and 510 mm, respectively. Two sites represent the different climate mesoregions. The terrains of both sites exhibit flat plateau surfaces where the mass movement was very weak. According to the age-depth relationship in the soil profile in this area (Zhou et al., 1994; Ding et al., 1999), a 2.0 m deep profile was dug from the surface on each site. Profile cutters were used to cut the profile so that each horizon was clearly visible (Fig. 1). The depth of each horizon, and soil profile characteristics of each horizon were recorded and photographed. The stratigraphic subdivisions and soil horizons were described based on color, texture, and structure of the two profiles in Table 1. The profile was divided into 40 layers with 5 cm thick depth increment. Bulk samples of 500 g for each layer were collected and sealed in bags.

2.2. Laboratory analyses

In the laboratory, after air-drying and picking out stone fragments and visible roots, bulk samples were passed through a 2-mm sieve. The ^{14}C chronology of 40 soil samples for each profile were measured in the Accelerator Mass Spectrometry Center, CAS Institute of Earth Environment. Bulk soil samples were oven dried (60 °C) to constant weight, then ground and sieved. In order to extract humic acids and to date the humin fraction alone, samples were first treated with acid and alkali solutions (Mook et al., 1983). All pre-treated samples were combusted by sealing the sample with CuO wire in an evacuated quartz tube, and then placing the tube in a 950 °C oven for 2 h. The resulting CO_2 was purified and reduced to graphite targets for AMS using the apparatus and methods described in Vogel et al. (1987). The ^{14}C dates obtained in the humin fraction should be considered as the minimum age of the SOM (Pessenda et al., 2001). The ^{14}C dates were calibrated using the methods of Reimer et al. (2013). The age-depth models were produced using a Bayesian model in the Bacon software (Blaauw and Christen, 2011).

Sub samples of the bulk soil were treated with H_2O_2 and HCl to remove organic matter, dispersed with $Na_4P_2O_7$ and treated with ultrasound for 3 min before particle size analysis. A standardized procedure was used for operating the Mastersizer 2000 (Malvern, UK) to measure soil particle size distributions. Additionally, the amounts of organic carbon were determined by a carbon analyzer with stepped heating routine (RC-412, LECO, USA). The calcium carbonate content was determined using a gasometric method (Glenn and Peter, 1983), and a MS2 Magnetic Susceptibility System (Bartington, UK) was used to measure magnetic susceptibility.

2.3. Reconstruction of past climate

The magnetic susceptibility of the Chinese loess was a good indicator of climate change in the Quaternary, especially variations in precipitation and temperature. Numerous studies have been carried out to correlate magnetic susceptibility with precipitation and temperature, and some equations were established (Maher et al., 1994; Liu et al., 1995; Han et al., 1996; Porter et al., 2001). These equations can be directly used to reconstruct paleoprecipitation and paleotemperature, if the soil has matured to, or close to, a magnetic steady state, in a few thousand years of active pedogenesis. The magnetic susceptibility of the Chinese loess is a rapidly formed soil property that has reached near-equilibrium with ambient climatic conditions (Maher et al., 1994). Therefore, the fitted equations using modern pedogenic susceptibility, precipitation, and temperature can be used to quantitatively reconstruct paleoprecipitation and paleotemperature on the Luochuan and Yanchang sites during the Holocene. The following two equations, which were constructed based on abundant sampling number in the Loess Plateau (Han et al., 1996), were chosen in our study:

$$MAP = -22.7 + 11.6\chi - 6.7 \times 10^{-2}\chi^2 + 1.9 \times 10^{-4}\chi^3 - 1.9 \times 10^{-7}\chi^4 \quad (n = 63,\ R^2 = 0.675) \tag{1}$$

$$MAT = -2.4 + 0.2\chi - 1.1 \times 10^{-3}\chi^2 + 2.7 \times 10^{-6}\chi^3 - 2.7 \times 10^{-9}\chi^4 \quad (n = 63,\ R^2 = 0.718) \tag{2}$$

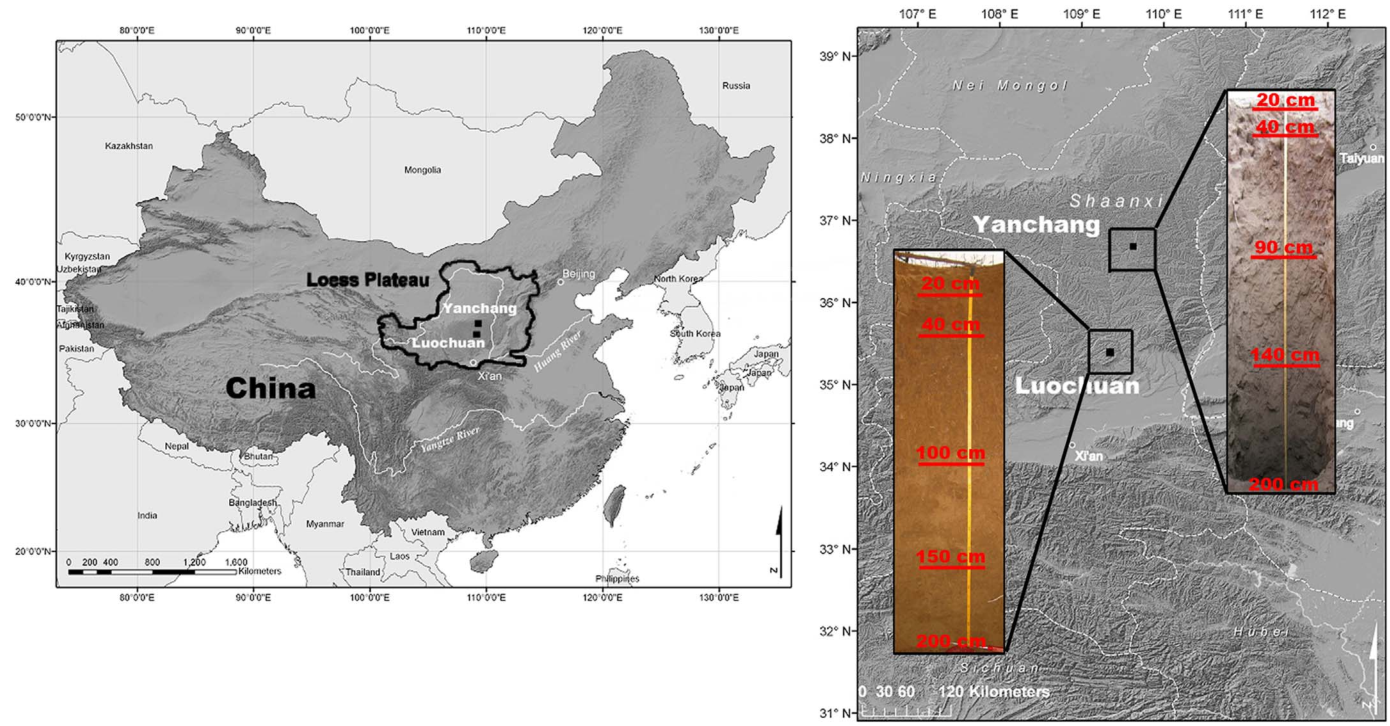

Fig. 1. Location of the Luochuan and Yanchang study sites in the central Loess Plateau of China, and photos of two profiles.

where MAP is mean annual precipitation (mm); MAT is mean annual temperature (°C); χ is magnetic susceptibility of soil (10^{-8} $m^3 \cdot kg^{-1}$).

2.4. Correlation equation of erosion and precipitation

The relationships between rainfall erosivity (R_e, MJ·cm·$ha^{-1} \cdot h^{-1} \cdot yr^{-1}$) and *NPP* and mean annual precipitation (P_m) in the Loess Plateau can be described by the following two equations (Xu, 2005):

$$R_e = 18.308e^{0.0036P_m} \quad (n = 152, \ R^2 = 0.591) \tag{3}$$

$$1/NPP = 10^9 P_m^{-3.6341} \quad (n = 283, \ R^2 = 0.801) \tag{4}$$

According to the relationships among soil erosion, precipitation, and vegetation (Xu, 2005), soil erosion was primarily determined by R_e and *NPP*, a non-linear relation would exist between soil erosion and precipitation. In our study, soil erosion intensity (I_e, t·$km^{-2} \cdot yr^{-1}$) which is defined as erosion amount per square kilometer per year and P_m were correlated using published data from 64 counties in the Loess Plateau of China in the mid-1980s (Xu, 2005). Before this period, the large scale and extensive soil and water conservation were not yet implemented, so the natural process of erosion had not yet been greatly changed (Wei et al., 2006).

3. Results

3.1. Stratigraphy, pedology, and chronology

The age-depth model was developed using a Bayesian model in Bacon software (Fig. 2). The model shows four distinct phases of soil accumulation in two profiles. The fastest accumulation rate, 24.0 and 20.0 cm·kyr^{-1}, were in S_{01} layer in the Luochuan and Yanchang profile, respectively. Both profiles had the same accumulation rate, 13.3 and 16.7 cm·kyr^{-1}, in top two layers (TS and L_0) and S_{02} layer, respectively. But the accumulation rate was very different in L_1 layer, 12.5 and 17.1 cm·kyr^{-1} in Luochuan and Yanchang, respectively.

The stratigraphic subdivisions and soil horizons were described based on color, texture, and structure of the two profiles in Luochuan and Yanchang (Fig. 3). The Holocene sequences overlaid the Malan Loess (L_1) layer of the last glaciation in both profiles. The presently ploughed topsoil (TS) was the upper part of the recent loess (L_0). The pale brown soil (S_0) was a complex soil (S_{01} and S_{02}) buried by the recent loess (L_0) and it was thus defined as a paleosol of the Holocene (Zhu et al., 1983). Calcium carbonate concretions (ca. 0.5–1 cm in diameter) were visible in the lower paleosol (S_{02}).

3.2. Holocene climate changes

Fig. 3 shows the changes of four substitutive climatic proxies, namely, magnetic susceptibility (χ), calcium carbonate content ($CaCO_3$), total content of organic carbon (TOC) and clay ($<$ 0.005 mm), in the two profiles of Luochuan and Yanchang. Both χ and TOC were reliable indicators for paleoclimate change, including both variations of paleoprecipitation and paleotemperature (Maher and Thompson, 1995; Jobbágy and Jackson, 2000; Liu et al., 2013). Magnetic susceptibility represented the concentration of ultrafine ferromagnetic grains produced in the topsoil during bio-pedogenesis (Maher, 1998). Total organic carbon reflected the enrichment of humus in soils (Balesdent et al., 1988). Therefore, these properties recorded the changeable intensity of pedogenesis which was influenced by both precipitation and temperature (Liu et al., 2013). The lowest values of both χ (36.77–51.15 × 10^{-8} $m^3 \cdot kg^{-1}$ and 36.20–41.59 × 10^{-8} $m^3 \cdot kg^{-1}$) and TOC (0.19–0.30% and 0.13–0.15%) were found in the Malan Loess (L_1) before 8500 cal. yr BP. These properties gradually increased in paleosol (S_0) from 8500 to 3000 cal. yr BP. The peak values of both χ (100.36–148.88 × 10^{-8} $m^3 \cdot kg^{-1}$ and 53.82–94.71 × 10^{-8} $m^3 \cdot kg^{-1}$) and TOC (0.41–0.59% and 0.26–0.47%) were present in the upper paleosol (S_{01}), hereafter they decreased in both recent loess (L_0) and ploughed topsoil (TS) from 3000 cal. yr BP to present. The changes of χ and TOC indicated that the pedogenesis and the contemporaneous precipitation and temperature were very weak before 8500 cal. yr BP, then gradually increased from 8500 cal. yr BP to a peak in 3000 cal. yr BP, and finally decreased during the last 3000 years. Furthermore, the values of both χ and TOC were larger on the Luochuan site than on the Yanchang site in the same period that showed a stronger pedogenesis, namely more precipitation and higher temperature.

Table 1
Stratigraphic features and soil horizon designations of soil profiles on the Luochuan site and Yanchang site on the central Loess Plateau, China.

Profiles	Soil horizons	Depth (cm)	Stratigraphic subdivisions	Pedological description	Calendar age (yr BP)
Luochuan site	A_{p1}	0–20	Top soil (TS)	Light grayish brown, 10YR7/6 (moist), clay loam, weak medium granular structure, slightly hard (dry), abundant roots, some carbonate pseudomycelia in small spots, gradual smooth boundary.	1700–0
	A_{p2}	20–40	Recent loess (L_0)	Color heterogeneity, darkish brown blocks in pale brown soils, clay loam, fine subangular blocky structure, hard (dry), abundant roots, some pseudomycelia in root channels, a few krotowinas filled with C-material, smooth boundary.	3000–1700
	B_t	40–100	Upper paleosol (S_{01})	Darkish brown, 7.5YR3/3 (moist), clay loam, fine subangular blocks break to granules, slightly hard (dry), few roots, plenty of pseudomycelia, smooth boundary.	5500–3000
	B_k	100–150	Lower paleosol (S_{02})	Very pale brown, 7.5YR5/3 (moist), clay loam, massive, hard (dry), some very fine pores, few pseudomycelia, some lime nodules of 0.5–1 cm in diameter, few roots, clear boundary.	8500–5500
	C	150–200	Malan Loess (L_1)	Yellowish brown, 10YR8/4 (moist), clay loam, blocks break to granules, plenty of pseudomycelia, slightly hard (dry), few roots.	12,500–8500
Yanchang site	A_{p1}	0–20	Top soil (TS)	Light grayish brown, 10YR7/6 (moist), loam, weak medium granular structure, slightly hard (dry), abundant roots, some carbonate pseudomycelia in small spots, gradual smooth boundary.	1700–0
	A_{p2}	20–40	Recent loess (L_0)	Grayish brown, 10YR6/6 (moist), clay loam, fine subangular blocky structure, hard (dry), abundant roots, some pseudomycelia in root channels, a few krotowinas filled with C-material, smooth boundary.	3000–1700
	B_t	40–90	Upper paleosol (S_{01})	Darkish brown, 7.5YR3/3 (moist), clay loam, massive, hard (dry), little roots, some pseudomycelia in root channels, a few krotowinas filled with C-material, smooth boundary.	5500–3000
	B_k	90–140	Lower paleosol (S_{02})	Grayish yellow, 2.5Y7/6 (moist), clay loam, massive, hard (dry), few pseudomycelia, some lime nodules of 0.5–1 cm in diameter, few roots, clear boundary.	8500–5500
	C	140–200	Malan Loess (L_1)	Light yellow, 2.5Y8/6 (moist), clay loam, massive, plenty of pseudomycelia, hard (dry).	12,000–8500

As one of climate indicators, $CaCO_3$ content mainly demonstrated weathering and leaching during pedogenesis that were connected with precipitation. $CaCO_3$ exists as calcite in the forms of inherited calcite minerals, secondary crystallite (pseudomycelia) and concretions in the loess-paleosol sequence over the Loess Plateau (Zhao, 2000). The lowest values of $CaCO_3$ content (0.39–1.02% and 3.47–4.97%) were present in S_{01} of the two profiles, which indicated that the weathering and leaching of $CaCO_3$ during the formation of S_{01} were very strong. It should be noted that the abrupt decrease of $CaCO_3$ content in the connection of S_{01} and S_{02} showed the eluvial-illuvial depth. That was a reflection of strong precipitation during formation of the paleosol S_{02}, but not an abrupt change of climate (Zhao, 2000). The highest values of $CaCO_3$ (13.31–14.92% and 13.10–14.88%) occurred in the Malan Loess (L_1) that showed a dryer period during its formation. Higher $CaCO_3$ content (1.86–2.75% and 6.07–7.83%) in both TS and L_0 horizon were largely from the inherited calcite that was deposited recently by dust-falls (Liu, 1985). The $CaCO_3$ had not been depleted yet under semi-arid climatic conditions at the present. Clay (< 0.005 mm) content was a reflection of the intensity of bio-pedogenesis that was also closely related with precipitation brought on by the south-eastern monsoon (Lu and An, 1998). The largest clay content (25.74–28.23% and 15.37–20.32%) occurred in the paleosol S_{01} that indicated wet conditions and active bio-pedogenic processes during soil formation. Lower values in S_{02}, L_0, TS and the lowest values in L_1 were respectively observed, that showed dryer conditions and weaker bio-pedogenesis in these horizons.

3.3. Reconstruction of paleoprecipitation and paleotemperature

The mean annual precipitation and temperature were estimated on the Luochuan and Yanchang site in the Holocene using Eqs. (1) and (2) (Fig. 4a1, a2, b1, and b2). In both areas, the lowest precipitation (320–418 mm and 316–355 mm) and temperature (4.2–6.1 °C) were present before 8500 cal. yr BP. Then they gradually increased from 8500 cal. yr BP to the peak (728 mm and 612 mm, 12.8 °C and 10.2 °C) in 3000 cal. yr BP. Hereafter, they decreased from 3000 cal. yr BP to present. The estimated values of present (639 mm and 516 mm, 10.8 °C and 8.2 °C) were very close to the recorded mean annual precipitation (622 mm and 510 mm) and temperature (9.2 °C and 8.9 °C). The wetter and warmer climate in Luochuan than in Yanchang may be attributed to the lower latitudes and the stronger summer monsoon (Yao et al., 2000; Maher, 2008).

3.4. Evaluation of natural erosion

The following correlation equation was obtained by fitting the data of mean annual precipitation and soil erosion intensity from 64 counties in the Loess Plateau of China in the mid-1980s (Fig. 5):

$$I_e = 1/(3.594 \times 10^{-3} - 1.524 \times 10^{-5} P_m + 1.637 \times 10^{-8} P_m^2) \quad (n = 64,\ R^2 = 0.639) \tag{5}$$

This equation can explain 63.9% of the variance in soil erosion caused by precipitation. The *F* test of Eq. (5) is 53.98 larger than $F_{0.01\ (1,\ 62)} = 7.08$. Eq. (5) is statistically significant with 99% confidence.

In the Loess Plateau of China, spatial gradient in physico-geographical conditions is high but continuous. This area is covered by a thick loess mantle and the grain size of loess becomes coarser from southeast to northwest (Liu, 1985). In addition, the spatial gradient in erosion and specific sediment yield is also high. Thus, this provided a possibility to evaluate the past natural erosion using the method of spatial sequence instead of time successional sequence. Eqs. (3), (4) and (5) were used to calculate R_e, *NPP* and I_e in the Holocene in the central Loess Plateau (Fig. 4a3, a4, a5, b3, b4, and b5). The results indicated that in Luochuan area, when precipitation increased from 320 to 450 mm during 12,500 to 7500 cal. yr BP, R_e grew from 58.0 to 95.5 MJ·cm·ha^{-1}·h^{-1}·yr^{-1}, but erosion-resistance or coverage of

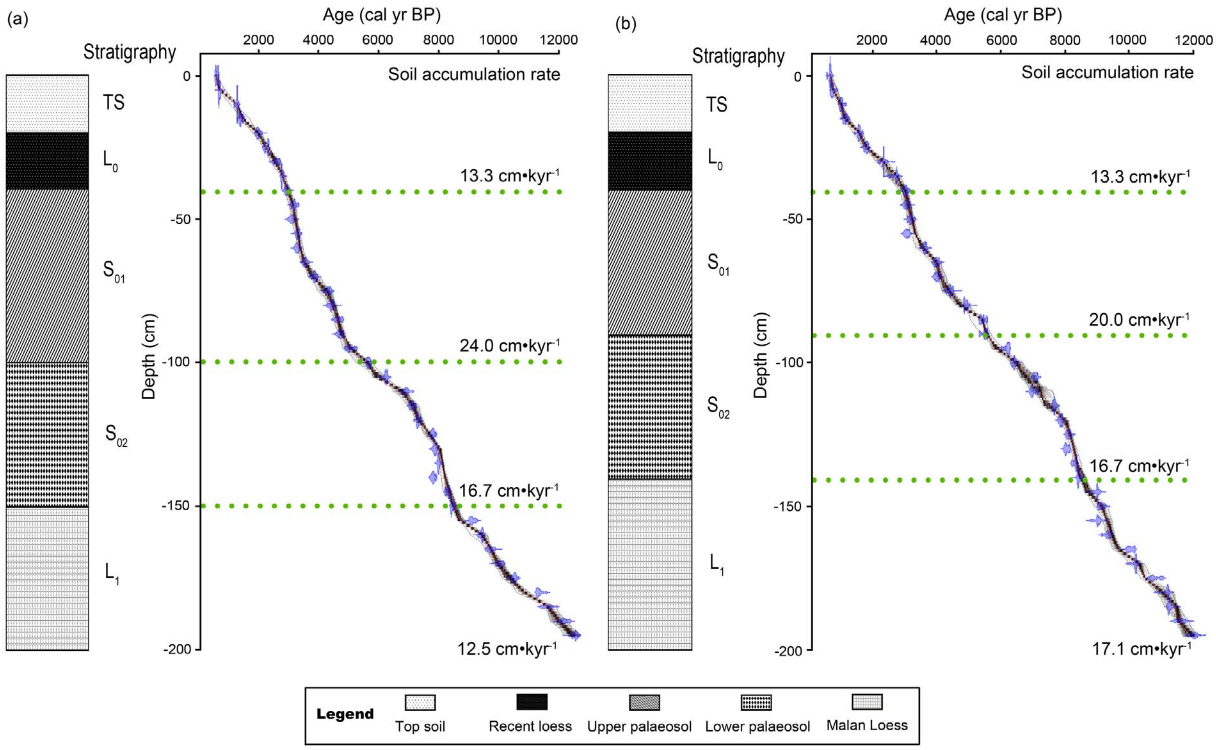

Fig. 2. Bayesian age-depth model for two profiles: (a) Luochuan site, (b) Yanchang site; which were produced using Bacon software (Blaauw and Christen, 2011). Calibrations were based on the IntCal13 curves (Reimer et al., 2013). Transparent blue shows the calibrated ^{14}C dates; darker grays indicate more likely calendar ages; gray stippled lines show 95% confidence intervals; red curve shows single best model based on the weighted mean age for each depth. The soil accumulation rates of different soil horizons are listed. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

vegetation was low (*NPP* $<$ 4.7 t·ha^{-1}·yr^{-1}). Around 5500 cal. yr BP, precipitation rose to another critical value of 530 mm. Although R_e increased 29.7% to 123.9 MJ·cm·ha^{-1}·h^{-1}·yr^{-1}, *NPP* sharply grew 72.3% to 8.1 t·ha^{-1}·yr^{-1}. Thus erosion intensity increased gradually from 2549 t·km^{-2}·yr^{-1} to a peak of 20,966 t·km^{-2}·yr^{-1}, and then decreased abruptly to 2052 t·km^{-2}·yr^{-1} before 5500 cal. yr BP. After that, erosion intensity maintained a low level and reduced with increasing precipitation due to higher resistance of vegetation (*NPP* $>$ 14.8 t·ha^{-1}·yr^{-1}). During the last 3000 years, an increasing trend of natural erosion could be observed because of reduced vegetation caused by decreased precipitation. The erosion intensity rose from 850 to 1850 t·km^{-2}·yr^{-1}. In the Yanchang area, natural erosion changed similarly. The precipitation increased gradually from 320 to 450 to 530 mm during 12,000 to 5500 to 3300 cal. yr BP, and that resulted in a change of erosion intensity from an increase to a decrease. However, the peak values of erosion intensity ranged between 12,703 and 21,148 t·km^{-2}·yr^{-1} and occurred during 5500 to 3300 cal. yr BP which is not in the half part of mid-Holocene. From 3300 to 1800 cal. yr BP, the precipitation increased higher than 530 mm, and *NPP* was larger than 12.3 t·ha^{-1}·yr^{-1}, so the erosion intensity decreased gradually from 3025 to 2809 t·km^{-2}·yr^{-1}. During the last 1800 cal. yr BP, *NPP* sharply reduced to $<$ 8.1 t·ha^{-1}·yr^{-1} when R_e was still high ($>$ 117.5 MJ·cm·ha^{-1}·h^{-1}·yr^{-1}). Thus the erosion intensity increased from 2809 to 11,177 t·km^{-2}·yr^{-1} quickly.

4. Discussion

4.1. Paleoclimatic evolution

Fig. 4 shows a gradually increasing trend of both MAP and MAT on both Luochuan and Yanchang site during the early Holocene. Liu et al. (2005) also found a recovery from the severe dry and cold glacial climate during Late Pleistocene and early Holocene in Luochuan. Then, the warm and wet climate was estimated from 8500 cal. yr BP to 3000 cal. yr BP (Fig. 4). This result also is supported by the independent evidence derived from other paleoclimate archives. Liu (1985) reported that Luochuan area was in warm and wet climate during mid-Holocene, because a large number of snail fossils named *Metodontia beresowskii* and *Metodontia hausaiensis* were found. Shi et al. (1993) reviewed the mid-Holocene climates based on palynology, paleobotany, paleozoology, archeology, paleopedology, paleolimnology, ice core, and sea level fluctuations since the 1970s, and concluded that the Megathermal in China was between 8500 and 3000 cal. yr BP. From 3000 cal. yr BP to present, the climate was cooling and drying according to the records of historical documents in China (Chu, 1973). This climate change was also recorded by the loess (Huang et al., 2002b; Tang and He, 2004). That cooling and drying trend is consistent with our results (Fig. 4). Therefore, the trend of our estimation on MAP and MAT can be confirmed.

Our results indicated that the Holocene climate change in both Luochuan and Yanchang showed temporal and spatial variation. However, the spatial variation of climate change in the Loess Plateau of China was rarely discussed. To the north of the Plateau, a sediment core collected from Lake Yanhaizi in the Inner Mongolia showed three humid phases, 13.4–8.0, 6.4–5.8, and 4.3–3.2 cal. kyr BP (Chen et al., 2003). In the north central Plateau, the Megathermal occurred from 7000 to 2700 cal. yr BP (Xiao et al., 2002) which started and ended 1500 and 300 years later than our results (8500–3000 cal. yr BP), respectively. The Megathermal was from 8500 to 3100 cal. yr BP (Huang et al., 2002a) in the southern Loess Plateau which was very similar to the central area. However, the warmest and wettest climate occurred in two periods of 8500–6000 cal. yr BP and 5000–3100 cal. yr BP, rather than in 5500–3000 cal. yr BP. The Megathermal was from 10,000 to 4000 cal. yr BP in the western Loess Plateau (Feng et al., 2006) which started and terminated 1500 and 1000 years earlier than the central region. Meanwhile, the warmest and wettest climate occurred between 7100 and 4490 yr BP which initiated and ended 1600 and 1490 years earlier than our results. Therefore, to describe the climate characteristics in the Loess Plateau of China, both temporal and spatial

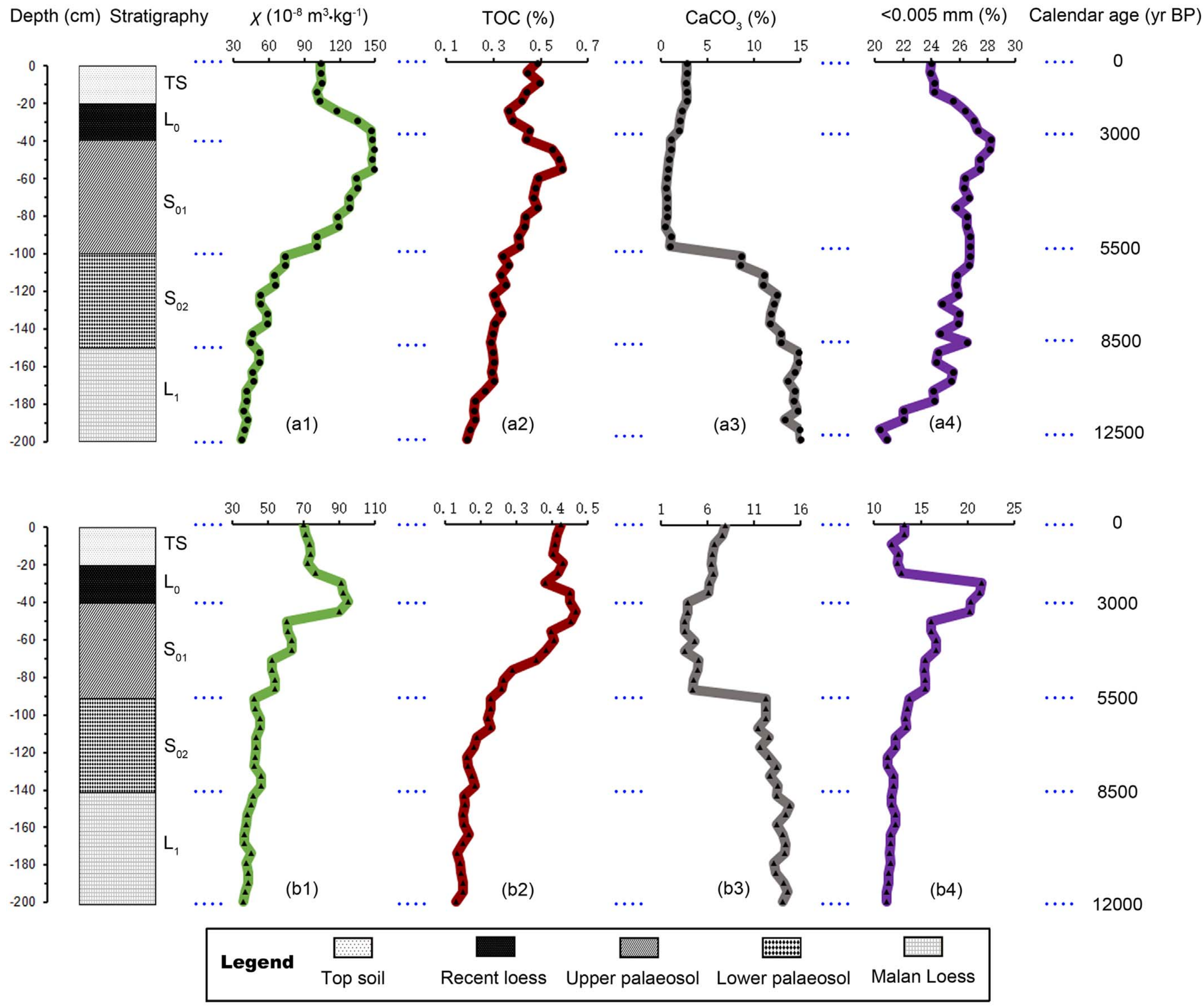

Fig. 3. Four substitutive climatic proxies, (1) magnetic susceptibility (χ), (2) total content of organic carbon (TOC), (3) calcium carbonate content ($CaCO_3$), and (4) clay (< 0.005 mm), change with time in two profiles of (a) Luochuan site and (b) Yanchang site.

variation should be noted. Especially, to summarize the general climate change in the whole area, spatial variation should be considered in detail.

4.2. Holocene erosion

Fig. 4 showed that the peak erosion values was 20,966 $t{\cdot}km^{-2}{\cdot}yr^{-1}$ in 7500 cal. yr BP and 21,148 $t{\cdot}km^{-2}{\cdot}yr^{-1}$ in 3300 cal. yr BP on Luochuan and Yanchang sites, respectively. These values didn't appear in neither warmest and wettest nor coldest and driest climate condition. This result was inconsistent with some research. Liu (1985), Yuan et al. (1987), and Deng and Yuan (2001) believed that severe soil erosion in Loess Plateau of China took place during humid-warm climate when the amount of precipitation was large. Jing and Li (1993) and Du and Zhao (2004) held an opposite point of view that the severe erosion coincided with the appearance of dry-cold climate, in which the vegetation coverage was low and rainstorms were still frequent despite less precipitation. However, the study of He et al. (2006) supported our results that soil erosion was more intense during the transition from dry-cool to wet-warm climate than during wet-warm and cool-dry climatic episodes.

The peak erosion may be caused by the combined effect of high frequency of rainstorms, loose loess, and low vegetation coverage. These results were contradictory, and that may be ascribed to three possible reasons. First, although most research noticed and analyzed the main factors of natural erosion, the analyses were qualitative. The relationships proposed between soil erosion, precipitation, and vegetation were speculative. It's very hard to accurately estimate the effect of precipitation and vegetation on soil erosion without a quantitative relationship among them, such as the Langbein-Schumm curve. Furthermore, the definition of the words to describe climate condition (e.g. wet, humid, warm, cool, cold and dry) and soil erosion (e.g. severe and weak) were also qualitative, which may have also contributed to the uncertain and conflicting results. Finally, some of studies tried to summarize the soil erosion change in the whole Loess Plateau of China whereas the asynchrony of Holocene climatic change and spatial variability in soil erosion within the entire area was ignored. The climate change was different from place to place, so severe erosion should not occur at the same time in all places.

Our study proposed a new approach to calculate historical soil erosion. Although the human activity-induced erosion has not been totally discounted from the data in this study, the estimated erosion intensity during the Holocene can show a principal trend of erosion

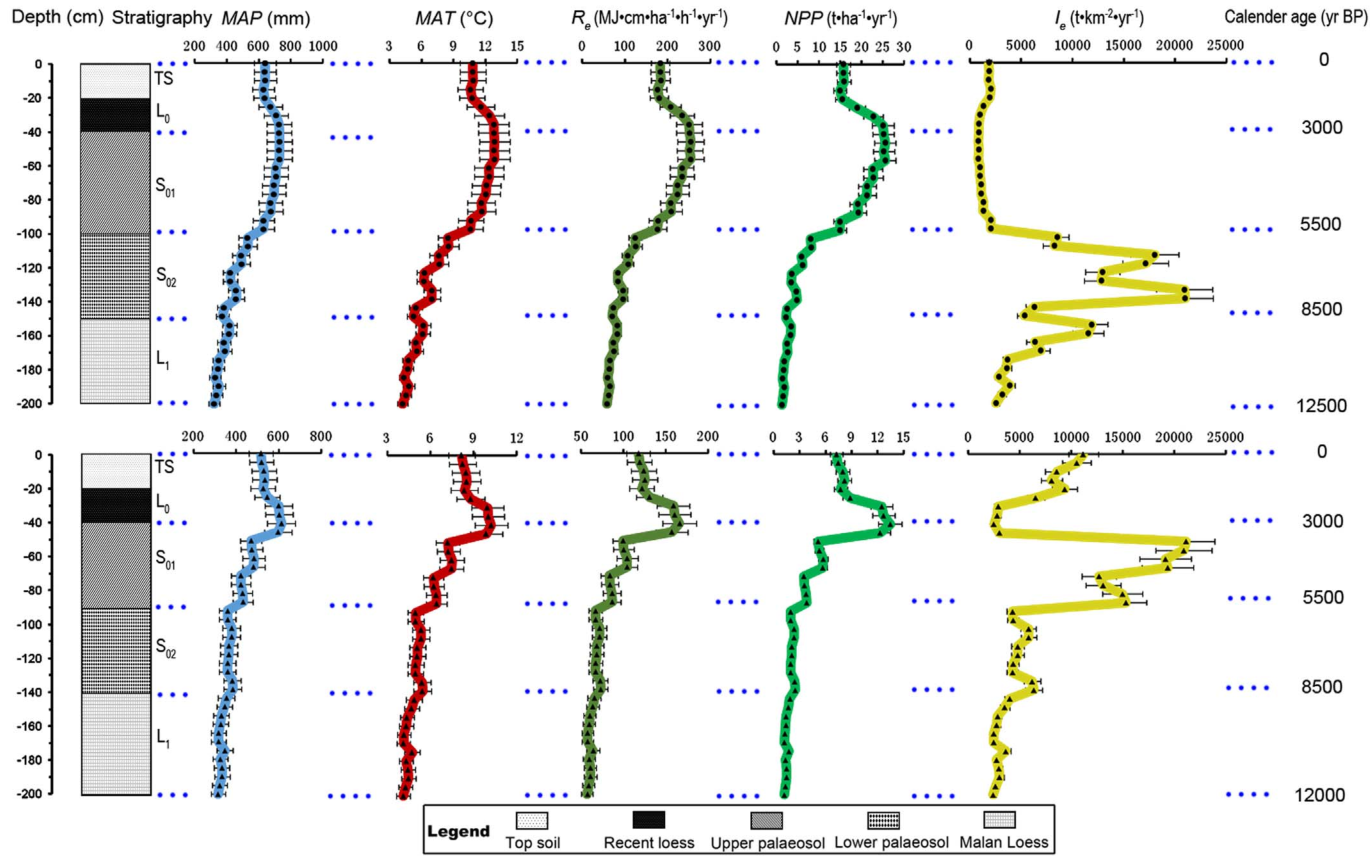

Fig. 4. Variation of calculated (1) mean annual paleoprecipitation (*MAP*), (2) mean annual paleotemperature (*MAT*), (3) rainfall erosivity (R_e), (4) net primary productivity (*NPP*), and (5) soil erosion intensity (I_e) during the Holocene in two profiles of (a) Luochuan site and (b) Yanchang site.

caused by precipitation. According to the Langbein-Schumm curve (Langbein and Schumm, 1958; Collins and Bras, 2008), human activities can only determine the absolute magnitude of soil loss. The natural erosion intensity may be overestimated due to human activity-accelerated erosion (Wei et al., 2006), but it was still a good indicator to show how natural erosion changed over long periods. It should be noticed that asynchronous Holocene climatic change would result in variation of natural erosion across the Loess Plateau of China. Our method also provided a possibility to compare erosion intensity between different areas in the past time. However, a limitation of this method is the difficulty to find perfectly preserved soil profiles without soil erosion or human disturbance, especially in severe erosion regions such as the central Loess Plateau of China.

5. Conclusions

Two profiles were investigated on the Luochuan and Yanchang sites in the central Loess Plateau of China. Four climate indicators, i.e.

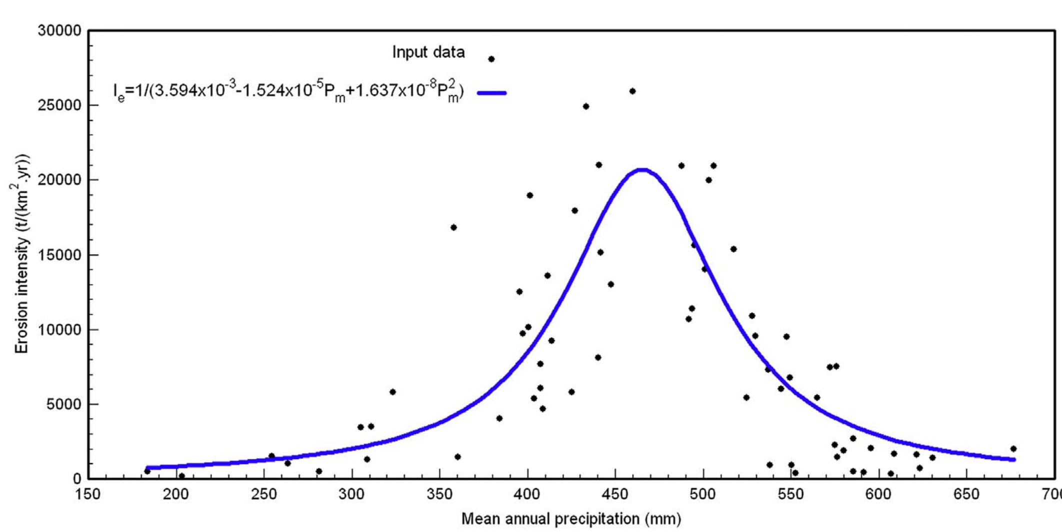

Fig. 5. Soil erosion intensity (I_e) as a function of mean annual precipitation (P_m). The fitting data of 64 counties were from a natural resources investigation in the Loess Plateau of China in the mid-1980s (Xu, 2005).

magnetic susceptibility, calcium carbonate content, total content of organic carbon, and clay, demonstrated the consistent trends of climate change during the Holocene. The climate was coldest and driest between 12,000 and 8500 cal. yr BP, and became warmer and wetter during 8500 to 5500 cal. yr BP. The warmest and wettest climate was from 5500 to 3000 cal. yr BP and was getting colder and dryer over the last 3000 years. Both mean annual precipitation and temperature were also estimated according to the fitted equations between modern pedogenic susceptibility and precipitation and temperature.

The relationship between soil erosion intensity and precipitation was established. It was used to calculate historical soil erosion intensity when combined with the estimated precipitation. The change of net primary productivity and rainfall erosivity were computed to interpret variation of natural soil erosion as well. The results indicated that, in Luochuan, erosion intensity increased gradually from 2549 $t{\cdot}km^{-2}{\cdot}yr^{-1}$ in 12,500 cal. yr BP to a peak of 20,966 $t{\cdot}km^{-2}{\cdot}yr^{-1}$ in 7500 cal. yr BP, and then decreased abruptly to 2052 $t{\cdot}km^{-2}{\cdot}yr^{-1}$ before 5500 cal. yr BP. After that, it continually decreased to a minimum of 850 $t{\cdot}km^{-2}{\cdot}yr^{-1}$ around 3000 cal. yr BP, and finally increased to 1850 $t{\cdot}km^{-2}{\cdot}yr^{-1}$ in the present. However, the change of erosion intensity was different in Yanchang, and the peak values of erosion intensity ranged between 12,703 and 21,148 $t{\cdot}km^{-2}{\cdot}yr^{-1}$, appearing during 5500 to 3300 cal. yr BP. Furthermore, more severe soil erosion with a faster increase was predicted, ranging from 6547 to 11,177 $t{\cdot}km^{-2}{\cdot}yr^{-1}$, during the last 1800 years.

This work developed a new method to quantify historical soil erosion triggered by climate change. This method not only can derive detailed soil erosion intensity changes with variation of climate, but also provide a way to compare soil losses between different areas.

Acknowledgements

We thank editors and two anonymous reviewers for their valuable comments on the manuscript, and thank Dr. Glenn Wilson and Dr. Xunchang John Zhang for their suggestions and improving the English writing of the manuscript. This research was jointly supported by the Innovative Talents Promotion Plan in Shaanxi Province (No. 2017KJXX-83), the National Key Technology Research and Development Program of the Ministry of Science and Technology of China (No. 2015BAC01B03-03), Natural Science Basic Research Plan in Shaanxi Province of China (No. 2016JQ4017), the West Light Foundation of The Chinese Academy of Sciences (No. 2014-91), and the Special-Funds of Scientific Research Programs of State Key Laboratory of Soil Erosion and Dryland Farming on the Loess Plateau (No. A314021403-C2).

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