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Holocene wildfire history and human activity from high-resolution charcoal and elemental black carbon records in the Guanzhong Basin of the Loess Plateau, China

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ABSTRACT

High-resolution sedimentary charcoal and black carbon (BC, including char and soot) records from a loess–soil profile, combined with magnetic susceptibility, $\delta^{13}C$ of soil organic matter of analyses, pollen counts and other paleoenvironmental proxies reveal past fire patterns and landscape evolution over the past 12,000 years. Results from the analyses of charcoal and BC influx show that regional fire activity was high in the early and late Holocene, whereas fire was less frequent and pervasive in the middle Holocene. Locally, fires were infrequent near the study site until the Late Holocene. Soot and char analyses do not parallel changes in charcoal variability, and thus appear to reflect either a different aspect of fire activity or else these data are registering aspects of particle transportation and deposition in addition to fire characteristics.

The patterns in fire activity observed during the Holocene are consistent with variations in vegetation inferred from $\delta^{13}C$ values in soil organic matter, pollen counts, and paleoclimate proxies. Drier and colder-than-present conditions on the Loess Plateau occurred during the Lateglacial and early Holocene (12,000–8500 years BP), which likely enhanced regional fire activity across the *Artemisia* and *Gramineae*-dominated steppe landscape in the south of the Loess Plateau. Wetter and warmer-than-present conditions during the mid-Holocene (8500–3100 years BP), reduced fire episodes and promoted the development of mixed forest and forest-steppe. The distribution of C_4 plants and woodland expanded at this time from 40% to 60% cover. The subsequent increase in fire-episode frequency during the past 3100 years is consistent with cooler and drier conditions in the late Holocene and also with changes in the spatial and temporal distributions of Neolithic burning practices, such as land reclamation and crop cultivation, during those periods (e.g., expansion of C_3 plants).

The close association between millennial-scale variations in fire and monsoon activity on the Loess Plateau suggests that future shifts in monsoon-related climate variability could have important consequences for fire and human activities as both respond to regional climate change.

1. Introduction

Spatial and temporal patterns of fire extent and frequency on semi-arid and arid regions are highly susceptible to seasonal climate changes (wet or dry), which affect both the vegetation type and fuel characteristics. On very long time scales (millennial), wildfire extent and frequency changes mirror trends in aridity in these ecosystems (Daniau et al., 2010; Pechony and Shindell, 2010). In addition, human land use has altered the type and distribution of biomass (fuels) on the landscape in the Loess Plateau for thousands of years (Huang et al., 2006; Tan et al., 2011). Over recent decades fire history studies from North America, as well as South America, Australia, central and boreal Europe, and temperate Europe have progressively contributed to our current understanding of Holocene fire–climate–vegetation–human interactions (Miller et al.,

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2005; Whitlock et al., 2007; Power et al., 2008; Vannière et al., 2008; Marlon et al., 2009; Kaal et al., 2011; Han et al., 2012). Recent studies of fire history in the Loess Plateau have mostly discussed the relationship between fires and climate changes on orbital to millennial timescales (Yang et al., 2001; Wang et al., 2005, 2012; Zhou et al., 2007). There are few high resolution or multi-disciplinary studies on Holocene fire history and human activity from this area, however (Huang et al., 2006; Jiang et al., 2008; Li et al., 2009; Tan et al., 2011).

Recent research on Holocene wildfire history and climate change in the middle reaches of the Yellow River of China show that the evolution of fire history across the region is closely related (though not linearly) to gradients in humidity, as well as to the build-up of burnable biomass and the spatial and temporal distributions of human activity (Tan et al., 2013). Such results are consistent the most important effects of climate on fire identified elsewhere, and include changes in temperature and precipitation that govern net primary productivity, and the abundance, composition, and structure of fuels (Carmona-Moreno et al., 2005; Moritz et al., 2012). Past and present fire patterns from the monsoon region, however, are quite different from inland parts of the Eurasian continent and central Asia in the westerly-dominated region (Huang et al., 2006; Umbanhowar et al., 2009). In the loess plateau, which is dominated by a monsoon climate, fire occurrence is attributed to fuel moisture, which is driven by seasonal precipitation (Tan et al., 2011, 2013). In northwestern Mongolia, however, temperature may be an important driver of fire activity, with minimum temperatures limiting biomass production and thus fire spread (Umbanhowar et al., 2009). The fire patterns observed in the Loess Plateau area are thus complex, and may reflect real inter-regional differences, contrasting sensitivities of different proxies, and/or differences in the degree of human versus climatic influences on vegetation change and thus fuels availability (Wang et al., 2013).

The complexity of the controls on regional wildfire patterns makes it impossible to evaluate different forcing factors (temperature/precipitation/vegetation) solely with site-specific single-proxy fire data. Thin section and sieving techniques for charcoal analysis, for example, have proven to be a robust proxy for local fire history reconstruction. However, such methods emphasize the importance of the identification and statistics of larger-sized charcoal that do not travel long distances (Clark, 1988), whereas they generally ignore smaller particles(< 10 μm) derived from regional biomass burning. This means that smaller charcoal particles, which may have been transported from extra-regional sources, are typically excluded from analysis. Restricting analyses to large particles, however, makes it difficult to estimate past regional fire frequencies, since most high-resolution charcoal records reflect local fires (within tens of kilometers). Charcoal syntheses, then, are typically characterized by high inter-site variability (Gavin et al., 2006). Furthermore, there are large uncertainties in identifying and quantifying changing fire regimes based on composites of local records, or of compositing lacustrine with marine sedimentary records, for example, which are based on very different analytical techniques (Chameides and Bergin, 2002).

Multi-proxy approaches to fire-history reconstruction are well-suited to examining the controls on fire in part because potential forcings vary across multiple temporal and spatial scales (Kehrwald, 2013). Charcoal can be used to reconstruct local fires, for example, while black carbon can be used to infer trends in broader-scale burning (Thevenon et al., 2010; Wang et al., 2013). Black carbon(BC), also referred to as elemental carbon (EC), is usually defined as the highly condensed carbonaceous continuum from the incomplete combustion processes of fossil fuels and vegetation fires (Goldberg, 1985). BC comprises two major categories, EC-char/charcoal and EC-soot/graphite particles, with no general agreement on how to define the boundaries between the two categories (Seiler and Crutzen, 1980). EC-char/charcoal is a solid residue at low temperatures, in which the cellular structure of plants are recognizable. EC-char/charcoal particle sizes are in the millimeter to micron range (particles generally >10 μm are called charcoal, Fig. 1). In contrast, EC-soot/graphite is characterized as nested graphitic-like spherical annuli, typically present in combustion-generated aerosol particles in the submicron range (Masiello, 2004).

Analyses of the particle size distributions of macro-char/charcoal particles (>50 μm) show that such particles are usually not transported far from fires, (Patterson et al., 1987; Clark, 1988), and are likely suitable for reconstructing local fire events (mostly deriving from within a few hundred metres). Micro-char/charcoal (<50 μm) reflects more regional fire history since the smaller particles travel longer distances from the deposition site. Yet even micro charcoal is thought to derive primarily from within 20–100 km from the site (Clark and Patterson, 1997). Submicron soot particles (<1 μm), however, can remain suspended in the atmosphere on the order of months and should therefore reflect the history of burning at a regional and global scope (Ogren and Charlson, 1983).

At present, black carbon (BC) concentrations can be measured by a chemo-thermal methods (CTO-375) (Gustafsson et al., 2001), thermal optical reflectance (TOR; Chow et al., 1993; Cavalli et al., 2010) and related methods (Kaal et al., 2008a, b). Comparison of these methods suggests that the TOR method as specified by the Interagency Monitoring of Protected Visual Environments (IMPROVE) protocol can provide strong differentiation between char and soot (Han et al., 2007). Analyzing patterns of charcoal and black carbon together, then, may allow us to shrink the spatial and temporal gaps in fire history reconstructions that could enable new insights into the global carbon cycle.

Located in the southern area of the Loess Plateau, Guanzhong Basin is a climate-sensitive region with the semi-arid and sub-humid Asian monsoon belt. The Basin also lies in the south-western part of the ecotone between traditional Chinese dry farming and nomadic pastoral practice. The region has experienced a long and complex land-use history, in which slash-and-burn cultivation via human-set fires and deforestation for dry farming has been occurring since at least 8000 years ago. Many Neolithic cultures that practiced dry farming with fire have since arisen on the Loess Plateau during the Holocene, including the Laoguantai Culture (^{14}C cal.7800–7000 years BP) and the Yangshao Neolithic Culture (^{14}C cal.7000–5500 years BP).

This study aims to reconstruct Holocene fire history at regional scales using a multi-proxy approach and in the context of other existing sedimentary charcoal and black carbon records. We then consider the history of paleofires in conjunction with archaeological data, and with available climate data, to investigate the role of monsoon-driven climate changes, vegetation dynamics and anthropogenic impacts on fire.

2. Regional setting and loess profiles

The Liangjiacun site (LJC) (34°28′N 107°53′E, 645 m asl), located on the Guanzhong Basin in the southern part of the Loess Plateau, is ca 125 km to the west of Xi'an and is adjacent to the eastern boundary of the site of the Bronze Age city Qiyi of the predynastic Zhou (Huang et al., 2000, Fig. 2). A typical Holocene loess–paleosol sequence was located and selected through field observation at the study sites. The soil horizons and the stratigraphic boundaries in the profile are well delineated and clearly identified (Fig. 3). It is believed that disturbance from natural processes and human activities were minimal during dust accumulation and soil formation.

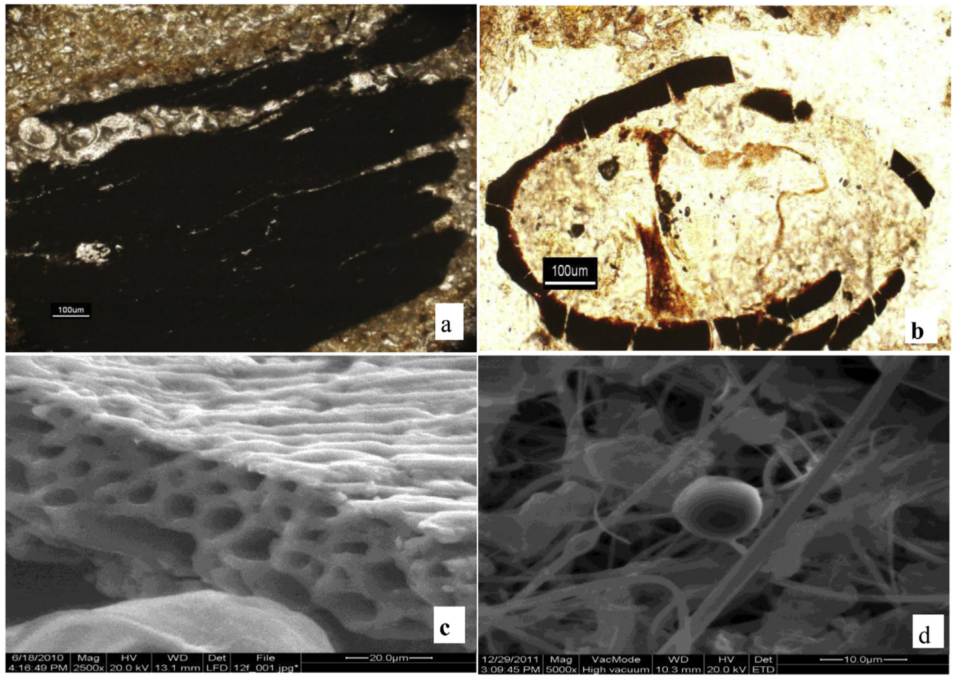


Fig. 1. Charcoal and black carbon microscopy images using plane polarized light and scanning electron microscopy (SEM) from the LJC site. (a) (b): charcoal microscopy image using plane polarized light (10 × 10); (c) charcoal on anatomical characteristics as revealed by SEM; (d) soot anatomical characteristics as revealed by SEM.

The mean annual temperature and the precipitation are 12 °C and 629 mm, respectively, with 50–70% of the precipitation occurring between July and September. It is warmer and wetter in summer, and colder and more arid in winter. Cinnamon -colored soils developed on the loess lands during the mid-Holocene, under mixed forest or forest steppe (Huang et al., 2009). Human cultural relics and settlements are commonly found here due to deforestation for cereal farming since the Bronze Age. Rain-fed cereal cultivation has been practiced in the area since 8000 years ago (early Neolithic times).

3. Sampling, stratigraphy, and chronology

After the two field investigations in the south of the Loess Plateau, a typical Holocene loess–paleosol sequence at the Liangjiacun (LJC) site was selected as the study site. The weathered surface of the cliff profile at the site was cleared first, then pedostratigraphic subdivisions were made through detailed observations of the color, texture, and structure of the sediments in the field (Fig. 3; Table 1). A total of 72 loess sediment samples were taken every 5 cm, spaced continuously down the study profiles

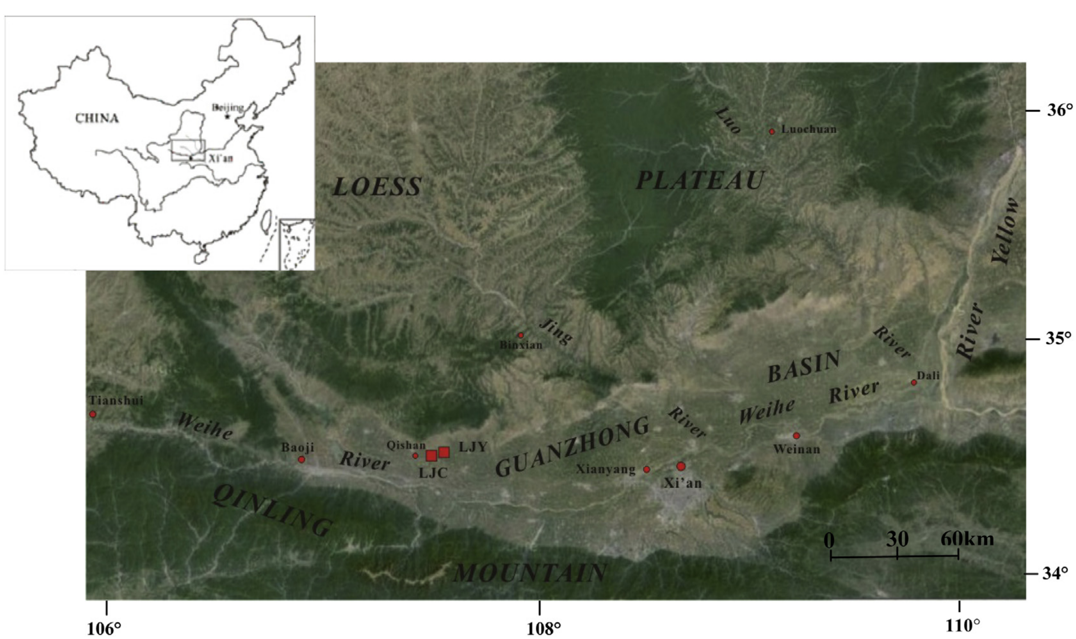


Fig. 2. Map showing the study region in the Guanzhong Basin in the southern part of the Loess Plateau. The study sites are marked as: Liangjiacun site (LJC), Liangjiayao site (LJY) (a previous study site, Tan et al., 2013).

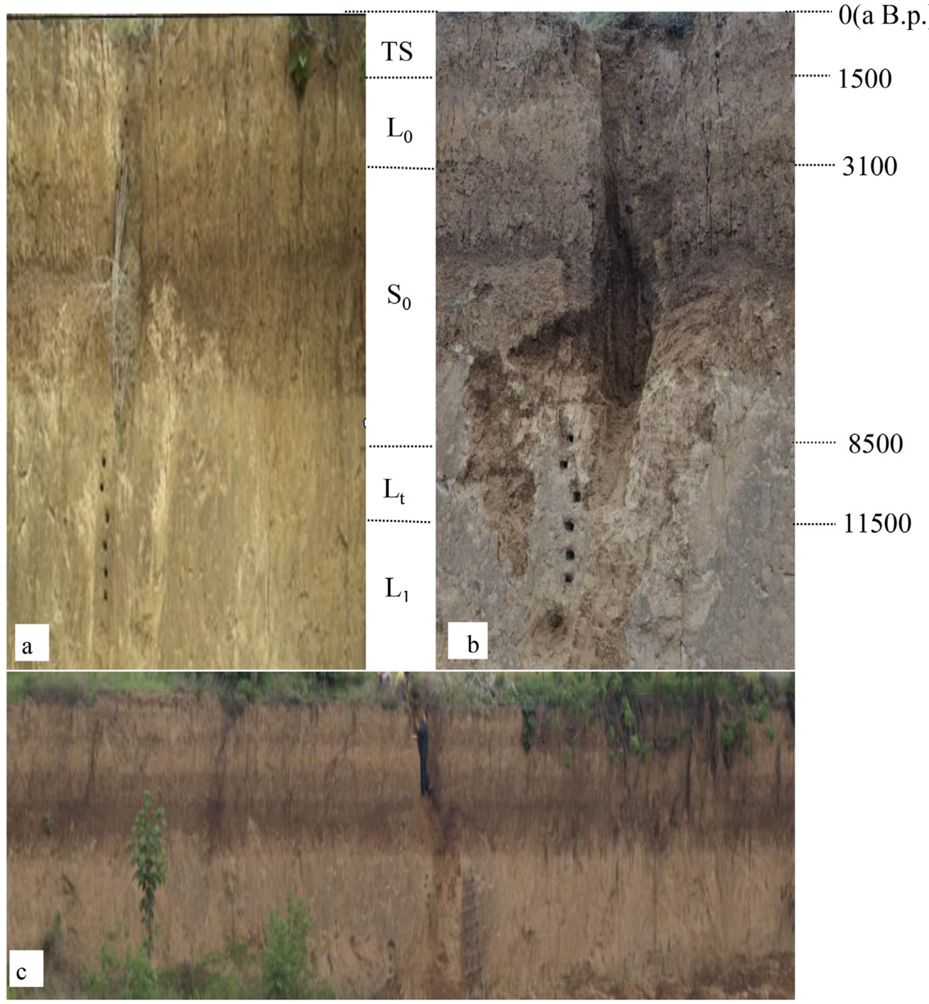


Fig. 3. Photographs of Holocene loess–paleosol profiles from the Zhouyuan loess tableland in the Guanzhong Basin in the southern part of the Loess Plateau: (a) the Liangjiacun site (LJC); (b) the Liangjiayao site (LJY); (c) photographs of the stratigraphic boundaries on the LJY and LJC profiles.

from the ground surface in the study site. The profiles at the LJC site and a further site at Liangjiayao (LJY, a previous study site; Tan et al., 2013) were located in a very flat or slightly shallow saucer-shaped area in the western Guanzhong Basin in the southern of the Loess Plateau. They are very close, about 100 m apart at the pedo-stratigraphy in the southern of loess tableland (Fig. 3). The profiles were exposed as a result of digging for brick-making by the villagers; they were well-preserved at the study sites and are pedo-stratigraphically well correlated with each other.

The chronological framework of the study profiles are based on ten OSL sediment dates from the LJY profile, stratigraphic correlations and archaeological identification of cultural remains retrieved from the profiles (Archaeological Institute of CASS, 1991, Table 2). The chronological framework established in the study sites is correlated well with other dated profiles in the Weihe River Basin (Huang et al., 2006; Tan et al., 2011, Fig. 4). An age–depth curve was established using the OSL sediment dates and archaeological identification of the cultural remains retrieved from the profiles (Fig. 5; Gansu Museum, 1960; Archaeological Institute of CASS, 1991; Tan et al., 2013).

The pedo-stratigraphic structure in the LJY and LJC profiles was typical of the Loess Plateau. In the two profiles, the pale yellowish orange colored, silty, friable, porous and carbonate rich loess (L_1), with the boundary between the Malan loess (L_1) of the last glaciation and the early Holocene transitional loess (L_t) was observed at a depth of 260 cm in the LJY profile and 320 cm in the LJC profile. An OSL date of 10,240 ± 500 years BP was obtained at a depth range of 270–265 cm in the LJY profile. This boundary is one of the most important stratigraphic marking-lines in the Chinese loess–paleosol sequence, indicating the end of the Lateglacial and the start of the Holocene at ca11,500 years BP (Roberts, 1992; Mayewski et al., 2004; Hoek and Bos, 2007). The slightly weathered or pedogenically-modified early Holocene transitional loess (L_t) was identified at a depth range of 260–210 cm in the LJY profile and 320–290 cm in the LJC profile. An Ustic Isohumisol (S_0) was identified at a depth range of 210–110 cm in the LJY profile and 290–110 cm in the LJC profile. An OSL date of 8440 ± 330 years BP was obtained at a depth range of 240–235 cm in the LJY profile. Anthropogenic remnants, including small fragments of gray pottery, were identified at a depth range of 120–110 cm in the upper part of the paleosol (S_0) in the LJC profile. These were identified as the remains of human activities during Pre-Zhou and Shang Dynasties (ca 3500–3100 years BP) (Tan, 2010). The Cinnamon soil (S_0) is widely distributed over the Loess Plateau and it was dated to 8500–3100 years BP (Huang et al., 2000). The soil (S_0) is underlain by the transitional loess (L_t) of the early Holocene, and it is buried

Table 1
A description of the paleo–soil profiles at the Liangjiayao site (LJY) and Liangjiacun (LJC) site on the Guanzhong Basin loess tableland in the southern part of the Loess Plateau.

LJY profile			LJC profile		
Depth (cm)	Stratigraphic subdivision	Pedological description	Depth (cm)	Stratigraphic subdivision	Pedological description
40–0	Topsoil (TS)	Pale orange, silt, medium granular structure, abundant bio-pores.	40–0	Topsoil (TS)	Pale orange, silt, medium granular structure, abundant bio-pores.
110–40	Recent loess(L_0)	Pale brown, silt massive structure,	110–40	Recent loess (L_0)	Pale brown, silt massive structure,
210–110	Paleosoil (S_0)	Red brown, clayey silt, granular structure; abundant bio-pores, well-rounded spherical, few earthworm burrows and excrement	290–110	Paleosoil (S_0)	Red brown, clayey silt, granular structure; abundant bio-pores, well-rounded spherical, few earthworm burrows and excrement, 120–110 cm, fragments of gray pottery
260–210	Transitional loess (L_t)	Pale yellow orange, silt, massive structure; some small $CaCO_3$ concretions	320–290	Transitional loess (L_t)	Pale yellow orange, silt, massive structure; some small $CaCO_3$ concretions
? –260	The Malan Loess (L_1)	Pale yellow orange, silt, constant massive structure,	? – 320	The Malan Loess (L_1)	Pale yellow orange, silt, constant massive structure,

by the recent loess (L_0) deposits and the topsoil (TS) of the late Holocene. The boundary between the soil (S_0) and the recent loess (L_0) was identified at a depth of 110 cm in the LJY and LJC profiles. It is the most widely distributed and easily identifiable pedostratigraphic marking-line in the Holocene loess–paleosol profiles in the middle reaches of the Yellow River (Liu, 1988). An OSL date of 3140 ± 230 years BP was obtained at a depth range of 120–115 cm in the LJY profile. The recent loess (L_0) and the topsoil (TS) of were identified above a depth of 110 cm during the past 3100 years in the study profiles (Figs. 4 and 5; Table 1).

4. Methods

Magnetic susceptibility was measured with an MS2 Magnetic Susceptibility Meter (0.47/4.7 kHz: Bartington, Witney, UK). The BC concentration from the LJC site was determined with a Model 2001 Thermal/Optical Carbon Analyzer (DRI, Reno, NV, USA) using the thermal/optical reflectance (TOR) method as specified by the IMPROVE protocol. All samples were pretreated with acid (HCl 18% and HF 48%) to remove carbonates and the mineral fraction as described in Han et al. (Han et al., 2007, 2009). The charcoal data used in this study was only from the LJY profile (Tan et al., 2013). Charcoal was quantified by counting in thin sections, as described in Tan et al. (2013).

According to the decomposition technique of Long et al. (1998), the BC (EC-char and EC-soot) influx (mg C/cm^2 per year) and macro/micro-charcoal influx (grain/cm^2 per year) were calculated by multiplying the BC (EC-char and EC-soot, mg/g) and charcoal (macro-charcoal >100 μm and micro-charcoal <25 μm, grains/cm^3) concentrations by the sedimentation rate (cm/year). BC (EC-char and EC-soot) and a time-series of the macro/micro-charcoal influx were then interpolated to pseudo-annual accumulation rates and binned in 50-year time intervals.

Table 2
OSL chronology data in LJY profile at the Liangjiayao site. Ten OSL samples and one charcoal sample were measured at the OSL Laboratory of Shaanxi Normal University in China (Tan et al., 2013).

Field no.	Lab no.	Depth (cm)	Material	Age (yr BP)
LJY-01	SNU-L001	27.5	Loess	1230 ± 230
LJY-02	SNU-L002	57.5	Loess	2080 ± 270
LJY-03	SNU-L003	87.5	Loess	2790 ± 320
LJY-04	SNU-L004	117.5	Loess	3140 ± 230
LJY-05	SNU-L005	147.5	Loess	3880 ± 340
LJY-06	SNU-L006	177.5	Loess	6770 ± 460
LJY-07	SNU-L007	207.5	Loess	7760 ± 470
LJY-08	SNU-L008	237.5	Loess	8440 ± 330
LJY-09	SNU-L009	267.5	Loess	10,240 ± 500
LJY-10	SNU-L010	297.5	Loess	11,900 ± 830

All the samples of $\delta^{13}C$ in soil organic matter were determined on a Finnigan-MAT-251 isotope ratio mass spectrometer (Thermo Fisher Scientific, Waltham, MA, USA) with a dual inlet system, and a CS-344 elemental analyzer (LECO, St Joseph, MI, USA). Sample isotope ratios were expressed as a permitted deviation relative to the Pee Dee Belemnite (PDB) standard, with a precision of <±0.2‰ or better (Liu et al., 2011). The relative abundance of C_4 plants as a percentage of the total vegetation biomass can be estimated using the equations proposed by Vidic and Montañez (Vidic and Montañez, 2004) for the Loess Plateau:% $C_4 = 100 * (27 + \delta^{13}C)/14$; where the average carbon isotope value of C_3 plants is −27‰ and that of C_4 plants is −14‰.

5. Results and interpretation

The variations in the micro-charcoal (<25 μm) flux indicate the changes in the regional occurrence of wildfire; macro-charcoal flux (>100 μm) indicates changes in local wildfire occurrence in the past (Whitlock and Larsen, 2001). In the LJY profile, higher values of the micro-charcoal flux occur in the recent loess (L_0) and Malan loess (L_1) layers, and lower values occur in the Luvisol/Isohumisol (S_0), indicating that regional fire activity was high in the early and late Holocene, whereas fire was less frequent and pervasive in the middle Holocene. Peaks in the macro-charcoal flux occur in the recent Loess (L_0), suggesting that local fires were infrequent near the study site until the Late Holocene.

Black carbon preserved in accretionary loess–soil profiles recorded changes in fire, which may be due to climatic variations, changes in vegetation, and human activities. EC-soot and EC-char influxes showed a stable rising trend since ca 8000 years, although a marked decline occurred in the past 3000 years. EC-char fluxes ranges from 0.003 to 0.078 mg cm^{-2} yr^{-1}; EC-soot fluxes ranges from 0.003 to 0.008 mg cm^{-2} yr^{-1} since 12,000 years BP. More specifically, EC-char and EC-soot values increased gradually from the early Holocene to relatively high at 3500 years BP and then sharply decrease. High-frequency variations in EC-soot and EC-char also show several shifts during the Holocene. Lower values occur prior to 8000 years BP, which may be attributed to fuel characteristics and low sedimentation rates (high deposition times) during this period. The increasing trends in the EC-soot and EC-char fluxes were relatively consistent from 7000 to 4000 years BP, and attained the highest value around 3500 years BP; an abrupt decline in overall BC values occurred subsequently.

Magnetic Susceptibility (MS) is the most widely used proxy to study Quaternary climatic change and the development of the loess–soil sequences on the Loess Plateau (Maher, 1998; Balsam et al., 2004; Huang et al., 2006). It was usually used to estimate the concentration of ultra-fine-grained ferromagnetic minerals,

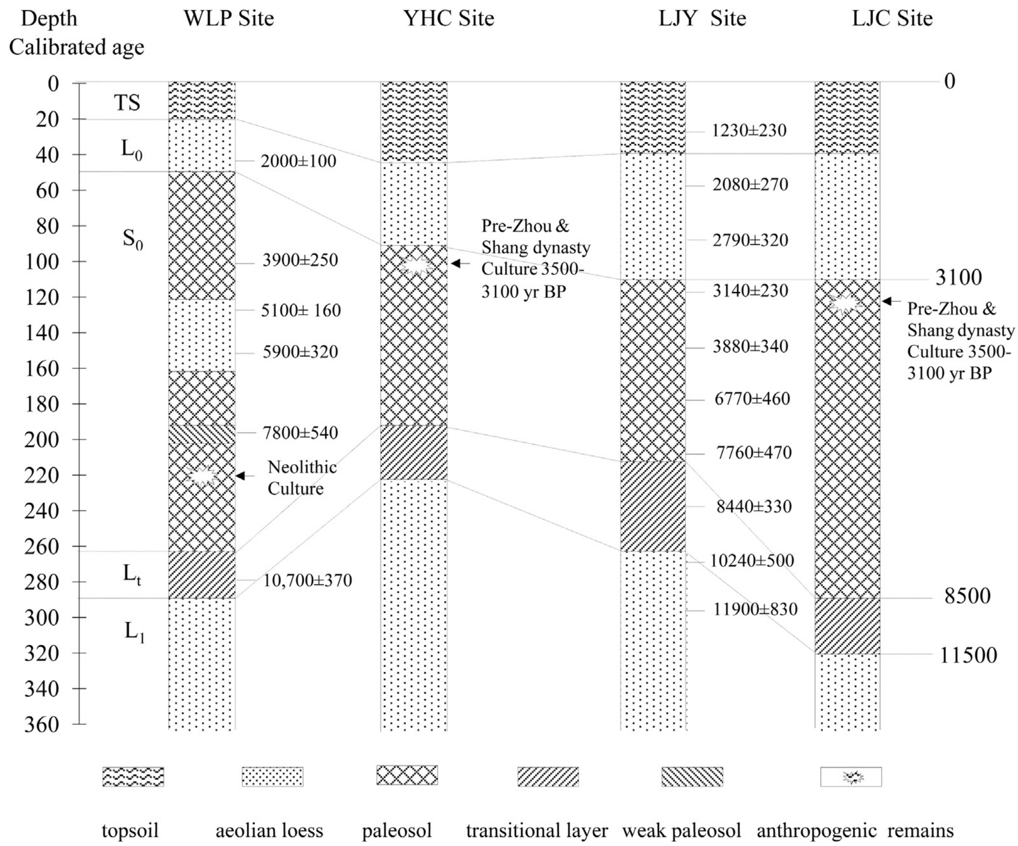


Fig. 4. Stratigraphic subdivision and chronology in the WLP, YHC, LJC and LJY profiles in the Guanzhong Basin in the southern part of the Loess Plateau.

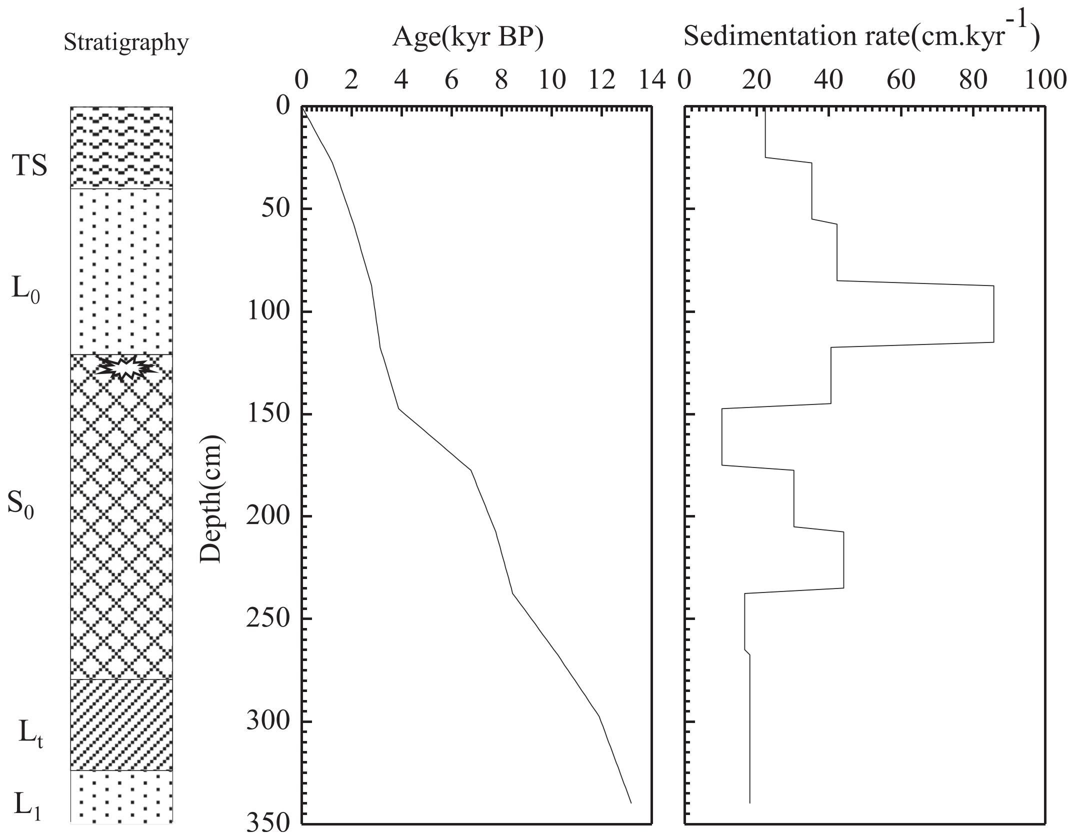
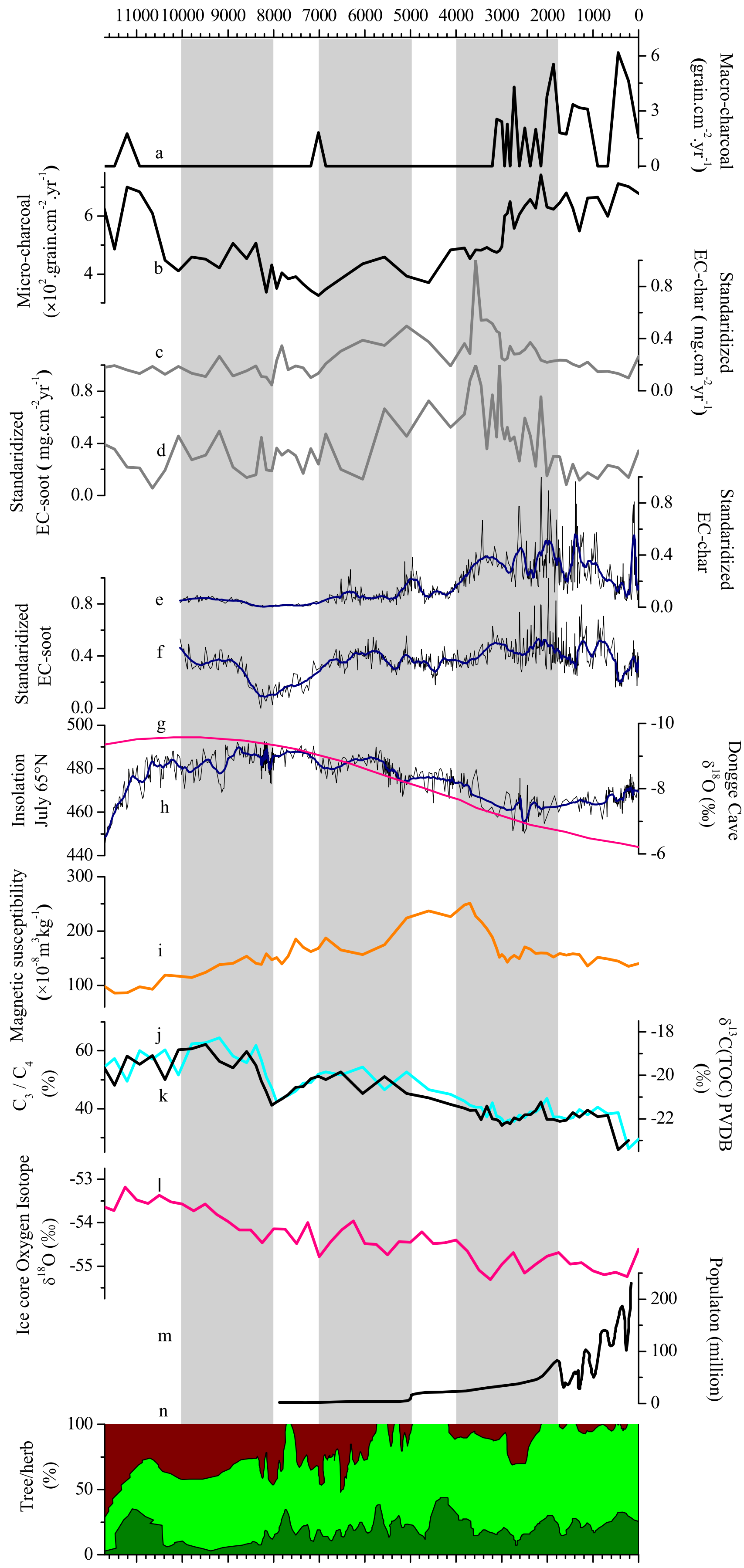


Fig. 5. Pedo-stratigraphic subdivision, the age/depth curve, and sedimentation rate of the LJC profile in the southern part of the Loess Plateau, marked by cultural remains from the Pre-Zhou and Shang Dynasties (ca 3500–3100 years BP).


Macro-charcoal (grain.cm-2.yr-1)
Micro-charcoal (×10^2.grain.cm-2.yr-1)
Standaridized EC-char (mg.cm-2.yr-1)
Standaridized EC-soot (mg.cm-2.yr-1)
Standaridized EC-char
Standaridized EC-soot
Insolation July 65°N
Dongge Cave δ18O (‰)
Magnetic susceptibility (×10-8m3kg-1)
C3 / C4 (%)
δ13C(TOC) PVDB (‰)
Ice core Oxygen Isotope δ18O (‰)
Populaton (million)
Tree/herb (%)
a
b
c
d
e
f
g
h
i
j
k
l
m
n
11000 10000 9000 8000 7000 6000 5000 4000 3000 2000 1000 0

which produced the topsoil (TS) during weathering and pedogenesis. MS records changes in the intensity of pedogenesis during dust accumulation, resulting from precipitation changes connected with monsoonal climatic variations (Maher, 1998, Maher and Alekseev, 2002). MS values vary between 52.5×10^{-8} and 251×10^{-8} m^3 kg^{-1} in the study profile. Lower values (52.5×10^{-8} and 140.3×10^{-8} m^3 kg^{-1}) occur in the Malan loess (L_1) of the Lateglacial period and in the transitional loess (L_t) of the early Holocene respectively, which indicates the very weak weathering alteration of accumulated dust during the last Lateglacial period. In contrast, high values (138.8×10^{-8} and 251×10^{-8} m^3 kg^{-1}) are observed in the paleosol (S_0), showing relatively concentrated ultra-fine-grained ferromagnetic minerals from pedogenesis during the Holocene Climatic Optimum. MS values decreased to below 130×10^{-8} m^3 kg^{-1} in the recent loess (L_0) and the topsoil(TS) indicating climate aridity and intensified human disturbance by arable cultivation. It should be noted that a sharp decrease in magnetic susceptibility values (below 160×10^{-8} m^3 kg^{-1}) are observed at a depths range of 200–185 cm in the LJC profile due to disturbances resulting from agricultural activities.

Soil organic matter and pedogenic carbonate in sediments have been widely used to reconstruct the proportion of plants using C_3 and C_4 photosynthetic pathways (Cerling, 1984; O'Leary, 1988; Cerling et al., 1989; Quade et al., 1989, 1995; Ehleringer and Cerling, 2002). The $\delta^{13}C$ record of organic matter is characterized by higher $\delta^{13}C$ values in the transitional layer (L_t) and the bottom of the paleosols (S_0), and lower values in the loess beds and the topsoil (TS). In the paleosol layer, the bottom values are higher than the upper values, implying more C_4 plants during the early Holocene than previously. The $\delta^{13}C$ of soil organic matter varies between −24‰ and −18‰ in the study profile. Lower values (−24‰) exist in the topsoil (TS) in the last few centuries (the percentage of C_4 plant < 20% cover). In contrast, high values (−18‰) are observed at the bottom of the paleosol (S_0) and transitional layer (L_t) during the early Holocene (the percentage of C_4 plant > 60% cover). There are two obvious shifts in the $\delta^{13}C$ of soil organic matter at intervals between ca 8000 years BP and 3000 years BP, which coincided with the change of C_4 plant percentages and were likely driven by seasonal precipitation variability during the shift from the early to mid-Holocene and from the mid-to late Holocene.

6. Discussion

Combining results from macro-charcoal, micro-charcoal and BC records can provide information about fire history at local, regional and broader scales. In eastern Asia, where there is a long history of relatively intensive human land-use, all types of paleofire data may be expected to reflect the importance not only of fire, but of changes in the seasonal monsoon and associated aridification during the Holocene, as well as shifts in cultivation and other human activities.

6.1. Fire, BC (char/soot) and Charcoal linkages

Macro/micro-charcoal and BC (EC-soot and EC-char) records from the study profiles show a generally increasing trend since ca 8000 years (Fig. 6 a–d). The timing of the increases, however, varies among the fire proxies. In the macro- and micro-charcoal data, peaks in the study profiles occur primarily during the late Holocene, after 3000 years BP. High values and the largest peaks are more prominent in the BC (EC-soot and EC-char) data during the mid-Holocene, however, from about 7000 to 4000 years BP (Fig. 6 a–d). The BC data from Lake Daihai, about 1000 km to the northeast of the southern Loess Plateau, shows trends more similar to the charcoal data from the study profiles than to the BC data, with the highest values registering during the late Holocene. The differences in the timing of the trends are not easily explained, and may be due to either real differences in fire history across spatial scales, or they may reflect differences in transportation mechanisms (Thevenon et al., 2010). Several aspects of the records can be interpreted in the context of the data on climate changes and human activities.

The records from BC (EC-char and EC-soot) and charcoal suggest changes in seasonal climate in relation to short-term climatic events (Fig. 6 a–d). There is a multi-fold increase in the amplitude of the dark carbon-rich particles between 3500 and 2800 years BP (Fig. 6 a–d), for example, which may correspond to climatic "deterioration" events (4200–4000 years BP) during the Holocene (Huang and Pang, 2002). Studies show that the mid-Holocene "Climatic Optimum" fell into a decline from ca 4500 years BP in China (Wang, 2005); Climate became highly variable and destabilized during the transition from the dominance of the maritime monsoon to the continental monsoon between 4500 and 3000 years BP (Wang and Gong, 2000; Xiao et al., 2004). Many lines of evidence from climatic proxies from China's monsoonal regions, have identified an abrupt climatic event between 4200 and 4000 years BP (L. Wang et al., 1999; Wu and Liu, 2004; Tan et al., 2008). High-resolution investigations of the cave speleothems, and lake and deep-sea sediments have documented a prominent climatic event between 4200 and 4000 years BP in other parts of the world. This event was associated with great droughts because of the reduced rainfall and river discharges (Cullen and deMenocal, 2000; Gasse, 2000; Bond et al., 2001; Yang and Williams, 2003; Booth et al., 2005; Arz et al., 2006; Parker and Goudie, 2008; Magny et al., 2009). High-resolution charcoal records and regional land-use history suggest that there was a population increase from the migration into the study region with abundant precipitation and heat as well as fertile soil around 4000 years BP. As a consequence, it led to the time of greatly increased biomass burning for intensified land reclamation for cereal cultivation via slash-and-burn agriculture. We can speculate that high levels of black carbon and charcoal influx were associated with human-caused deforestation and drought in the semi-arid to arid monsoonal regions.

Strikingly, the high-amplitude peak in the abundance of dark carbon-rich particles (3500–2800 years BP) also coincides with Pre-dynasty Zhou tribe activities associated with large-scale land reclamation for agriculture. Many Bronze-age cities were built on the Zhouyuan loess tableland in the western part of Ganzhong Basin during the Bronze Age and the early Iron Age (Huang and Su, 2009). Thus, many gray pottery fragments with one to two cm size(ca 3500–3100 years BP) from the Shang and Pre-dynasty Zhou periods were found at depths of 120 cm in the study profile

Fig. 6. Comparison of various proxies in the southern part of the Loess Plateau with other records during the last 12 ka; (a) macro-charcoal influx (>100 μm) at the Liangjiayao site (LJY) from the southern part of the Loess Plateau (Tan et al., 2013); (b) micro-charcoal influx (<25 μm) at the LJY site (Tan et al., 2013); (c) standardized EC-char influx values at the LJC site; (d) standardized EC-soot influx values at the LJC site; (e) standardized EC-char values from Lake Daihai (Han et al., 2012); (f) standardized EC-soot values from Lake Daihai (Han et al., 2012); (g) July 65°N isolation is marked by a red line (Berger and Loutre, 1991); (h) $\delta^{18}O$ data from Dongge Cave (Dykoski et al., 2005) indicating regional moisture variations; (i) low-frequency of magnetic susceptibility data at the LJC site; (j) value of $\delta^{13}C$ of soil organic matter data at the LJC site; (k) composite of C_3 and C_4 plants; (l) $\delta^{18}O$ data from the Dome Fuji Ice core (Kawamura et al., 2007) indicating the change of temperature in the northern hemisphere; (m) estimated population changes in China since 10 ka; (n) the tree/herb pollen (light green indicates herb; dark green indicates tree; brown indicates fern) from the southern part of the Loess Plateau (Han., 2000). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

(Institute of Archaeology, Archaeological Institute of CASS, 1991, Fig. 4). There is a minor decrease in the BC and charcoal fluxes at intervals of 8200 years (dated to OSL 8270–8043 years), and thus it is possible that the minimum in these data was synchronous with the 8200 year event. There may have been limited vegetation that inhibited fire spread, with cool and dry climate conditions registered by records of pollen, magnetic susceptibility, and $\delta^{18}O$ in an ice core from the Dome Fuji Ice Core (Antarctica) at ca8200 years BP (Fig. 6 i, l, n). Also, a decline at 8200 years BP is consistent with the EC-soot from Lake Daihai (Han et al., 2012). Another decrease in fire inferred from the declining BC and charcoal fluxes was between ca1500–1000 years BP. By the time of the Sui Dynasty (ca 581–618 A.D), all of the cultivable land areas had been expanded and the agricultural landscape had already been fully established at the study site (Tan et al., 2013), which may explain this decline in fire. Additionally, the late-Holocene BC decline may be attributed to effective moisture variability on century to millennial timescales and regional differences in land use (Tan et al., 2013). Such spatial heterogeneity in fire pattern is expected given the complex combination of climate changes and human activities and both respond to rapid climate variability (Marlon et al., 2013).

6.2. Fire and vegetation dynamics in response to monsoon climate change

Shifts in vegetation cover and composition associated with monsoon climate change may alter effective moisture. Climate change, which can induce increases in area burned, shifts in vegetation type, and changes in fire severity, are more likely to affect black carbon than charcoal emissions from biomass burning.

Recent studies infer that steppe and forest-steppe may exist in a “threshold ” state in which an herb-dominated ecosystem is maintained by frequent fires, while a tree-dominated (more than 40%) ecosystem has suppressed the spread of fire (Archibald et al., 2009). Previous studies show that fire occurred infrequently when precipitation was less than 300 mm in the arid steppe and more than 800 mm in humid broadleaf forests (Huber et al., 2004), While fire-conducive conditions remained within the range of precipitation between 300 and 800 mm during the Holocene. Charcoal (and BC) influx and the $\delta^{13}C$ of soil organic matter, along with pollen records, reveal a close relationship between vegetation-fire-climate in the study site. Three intervals are apparent in the data; the Lateglacial period and the early Holocene (12,000–8500 years BP), the mid-Holocene (85,00–3100 years BP), and the late Holocene (3100–0 years BP).

In the Lateglacial and the early Holocene periods, climate proxies from the Malan loess(L_1) and the transitional loess(L_t) at the LJC profile indicate that the climate was drier and colder than the present during those periods. Pollen records that show vegetation cover was characterized by an *Artemisia* and *Gramineae*-dominated steppe landscape (Han, 2000). Lower magnetic susceptibility values show that bio-pedogenic alteration of the accumulated dust was minimal, resulting from the lack of soil moisture and rainfall, as well as vegetation cover. Higher micro-char flux shows regional fire activity was high.

The evidence of high $\delta^{13}C$ in soil organic matter and the Dongge Cave $\delta^{18}O$ values, together with the increased magnetic susceptibility, suggests that a high seasonal precipitation coincided with high minimum temperatures during the growing season. Climatic amelioration thus resulted from the shifts from dry-cold glacial to humid-warm interglacial conditions in the Loess Plateau, which favored C_4 plant growth during the early Holocene (Sage et al., 1999). Herb biomass provides highly flammable fuel for fires. As with modern Mongolian arid steppe, natural fires frequently occurred with high eolian activity and an arid climate conditions in spring (Tan et al., 2011, 2013). At that time grassland fires likely tended to be of low intensity, injecting particles only into low atmospheric levels; this may have yielded a very limited elemental black carbon flux and a large charcoal influx to the deposition site (X. Wang et al., 1999), which were recorded as a high macro-charcoal influx and low BC influx during last Lateglacial period and the early Holocene.

Paleoecological records show that higher fire occurrence probably delayed the expansion of broadleaved evergreen forest by about 2000 years across the southern of the Loess Plateau (Tan et al., 2013). A subsequent increase in aridity until 8200 years BP resulted in the loss of much of the remaining C_3 woody vegetation, while increasing drought and/or fire activity in the future could favor the expansion of drought-adapted C_4 plants throughout the Loess Plateau. These results were attributed to regional seasonal precipitation variability linked with the instability of the South-eastern Asian monsoon during the early Holocene (An et al., 2000). Although, strong summer isolation intensified monsoons in the Northern Hemisphere during the early Holocene (Berger and Loutre, 1991), the dry climate (with low effective moisture for soils and vegetation) in the Loess Plateau during the early Holocene may have delayed the increase of sea-surface temperatures (SSTs) in the North Atlantic Ocean (Chen et al., 2008). In addition, there may have been enhanced subsidence of dry air masses induced by the maximum summer insolation in the early Holocene (Broccoli and Manabe, 1992; Zhao et al., 2007). Hydrologic changes in the American tropics and in Australasia are different from those in Asia and Africa and also from each other, but both show drier conditions in the early than late Holocene; these conditions are also paralleled by increasing fire before 8000 years BP (Marlon et al., 2013). In other words, lower effective moisture variability and fuel availability is likely the limiting factor for fire spread in the monsoon area during the early Holocene.

During the mid-Holocene (8500–3100 years BP), climate proxies from the Malan loess (L_1) and the transitional loess(L_t) in the LJC profile indicate that the climate was warmer and wetter than the present during those periods. The high magnetic susceptibility and negative $\delta^{18}O$ values from the Dome Fuji Ice Core show that bio-pedogenic alteration of the accumulated dust was very strong, resulting from the enhanced precipitation, soil moisture and vegetation cover with elevated temperatures during those periods. Pollen records show a decrease in C_3 vegetation followed by a fluctuating dominance of grasses C_4 and woodland vegetation (steppe dominated by *Artemisia* with some *Quercus* and *Ulmus*; Han, 2000), until the development of the mixed forest and forest-steppe with warmer and wetter during that period (Fig. 6 j, k, n). The climate favored the growth of C_4 plants and woodland vegetation cover (Obviously the increasing proportion of woody plants) began to expand from 40% to 60% with the increased warm-season precipitation that was induced by the enhanced the summer monsoon, as evidenced by the high value of $\delta^{13}C$ of soil organic matter record (Fig. 6 j, k; Wang et al., 2008). Fire activity is then greatly reduced during the period as registered by BC and charcoal fluxes (Fig. 6 a–d).

Such results are consistent with the gradual increasing trend in micro-sized charcoal particles and BC influx between in the LJC site and Daihai lake records, and suggest spatially coherent fire patterns of the change in relation to regional variations in the rapid variability of climate from the south of the Loess Plateau to Inner Mongolia (e.g.ca 6000–5000 years BP, Fig. 6 a–d). During this time, vegetation shifted from woodland toward steppe in association with cooler summers than before and weakened Asian monsoon. Effective moisture levels were slightly lower, and the greater proportion of woody plants fuels may have been more conducive to the spread of fire and produced abundant BC particles (Han et al., 2012), On the other hand, a high peak of macro-charcoal influx was identified in the paleosol (S_0) (7200 years BP; Fig. 6, b) at the study site, where many archaeological sites have been found within a radius of 3 km nearby (cal.7500–6500 years BP, Institute of Archaeology, Archaeological Institute of CASS, 1991). It is inferred that local fires must have been used for forest clearance for millet cultivation by the Neolithic people (cal. 7800–7000 years BP). The natural landscape did not change at all at the very beginning of dry farming, however, which implies that fire was controlled by effective moisture and its effects on fuel availability.

In the late Holocene, climate proxies from the Loess (L_0) and the top soil (TS) at the LJC site indicate that the climate became more arid. Aridification coincided with a widespread monsoonal decline since 3100 years BP. The decreased magnetic susceptibility shows that bio-pedogenic alteration of the accumulated dust was reduced, and dust accumulation was re-accelerated because of the decreased seasonal precipitation during those periods. Pollen records indicate that vegetation cover was characterized by a forest steppe landscape during those periods (Han, 2000, Fig. 6 n). Local biomass burning inferred from the macro-charcoal records appears to be very high during the last 3100 years (Fig. 6 a). Such results can be attributed to the spatial and temporal distribution and intensification of Pre-Zhou and Shang dynasties anthropogenic burning practices, such as deforestation, land reclamation, and crop cultivation during those periods (Tan et al., 2011).

Population records from the late Holocene indicate exponentially increasing trends since 3000 years. In north China, with the rapid population growth, human land-use for arable cultivation has become more intensified in the past 3000 years. Biomass burning and the cultivated area both increase sharply to 2000 years BP, suggesting that fires linked to the expansion of agriculture may have led to increased biomass burning in many regions at that time. A similar pattern of frequent fire occurrence during the late Holocene has been revealed by charcoal and BC records at Daihai Lake, which was attributed to widespread agriculture activity in the region (Huang et al., 2006; Wang et al., 2013). However, high micro/micro-charcoal and BC influx peaks with climatic variations as well, such as with the shift between dry and wet conditions associated with the Medieval Warm Period and Little Ice Age. Such parallels suggest that changes in monsoon intensity or effective moisture remain associated with fire activity even during the past 2000 years. Such results are also consistent with changes globally (Han et al., 2012; Marlon et al., 2013).

Judging from the charcoal and BC records and regional human land-use history of the past 3000 years, the most recent interval of frequent fires may reflect local variations in effective moisture as well as regional heterogeneity in ignition sources. Therefore, fire occurrences are attributed to the variability of seasonal precipitation and human land use intensification, as well as to regional monsoon-related climate changes.

6.3. *Fire and human disturbance in response to the climate variation*

Peaks in the charcoal and BC fluxes indicate a striking synchronicity not only with the major documented regional/local fire history (Tinner and Hu, 2003), but also the human land use and socio-cultural changes over periods of global climate cooling and regionally unstable climate conditions (Magny and Haas, 2004).

When the climate became warmer and wetter during the mid-Holocene, natural wildfires were greatly reduced as shown by the lower charcoal and BC influxes at the study site. However, a high-amplitude peaks registered by the macro-charcoal influx also coincides with changes in human practices for forest clearance for millet cultivation by human-set fires in the southern Loess Plateau. Both archeological and paleoecological data shows that many settlements of the Laoguantai culture were established between 7800 and 7000 cal years BP. Abundant rainfall controlled by the southeastern monsoon meant abudnant fertile lands in the southern Loess Plateau. As a result, the process of erosion and redeposition was intensified, as evidenced by markedly declined values of magnetic susceptibility at the study site (Huang et al., 2000; Tan et al., 2011, Fig. 6 i). Human beings thus apparently began to drive semi-arid ecological landscape succession, although the natural landscape did not change at all at the very beginning of dry farming (Tan et al., 2013).

A peculiar multi-fold increase of dark carbon-rich particles and charcoal influx appeared during the onset of a drier climate in the region during period of 3500–2800 years BP. Archeological and paleoecological records show that a number of settlements were built at this time and handicraft and bronze industries developed. The Pre-dynasty Zhou and Shang Dynasties were established on the Zhouyuan loess tableland since the Bronze Age (Huang and Su, 2009).

Although fire occurrence depends on a variety of natural climatic, vegetation or geomorphologic factors, there is a major contribution of humans to ignitions during the past 3100 years (Huang et al., 2006; Tan et al., 2011, 2013). Fires are used to clear or exploit the forests, to clear the land (i.e., control weeds, shrubs, tree seedlings, and litter accumulation), or to manage grazing land. Fires are also widely used for agricultural purposes (burning agricultural wastes and increasing the levels of nutrients available for uptake by plants). Fire is also used to produce charcoal for industrial and domestic uses (traditional metalworking or brick making, cooking and heating; Thevenon et al., 2010). The application of fire in land-use changes is especially widespread in the study regions as indicated by a number of population and land-use histories from the Loess Plateau covering the past 3100 years (Archaeological Institute of CASS, 1991).

During the climatic deterioration period that the Luvisol (S_0) overlapped with the recent loess accumulation layer (L_0) at the study sites in the late Holocene. Pre-dynasty Zhou tribial people in the middle of the Loess Plateau could not manage the land for dry farming as the consequence of drought and water shortage, and meanwhile they encountered the northern Nomadic tribe invaders in this region. Historical documents reveal that the ancestors of the Zhou tribe had to migrate to the south (the study sites), which was appropriate for dry farming development because of the better hydrothermal conditions. These prehistoric periods of human impacts in the southern of the Loess Plateau indicate a striking synchronicity not only with the major documented socio-cultural changes and the regional fire history (Tinner and Hu, 2003), but also with the periods of global climate cooling and weakened Asian monsoon (Magny and Haas, 2004). Therefore, it suggests that unstable environmental conditions influenced human settlement patterns in the southern of the Loess Plateau together with land-use changes or technological and social innovations (Huang and Pang, 2002).

6.4. *Conclusion*

The combination of BC and charcoal records in loess sediment with other paleoclimatic proxies provide us with unique information about past changes in fire history at local-to-regional scales. The results confirm that macro-charcoal data indicates local fires, whereas black carbon (char and soot) reflects regional fires. All the records show high fire occurrence during the late Holocene, and indicate strong relationships between climate changes and regional biomass burning. However, the possibility of differences in transportation mechanisms and biomass burning processes at regional and local scales for charcoal and BC requires further study. Two significant shifts in the abundance of C_4 plants were aligned with the evolution of the natural landscape from forest steppe to steppe, forced by changes in seasonal precipitation around the period of 8000 and 3000 years BP (Sun et al., 2012, Fig. 6 j, k). Such changes attest to the importance of the weakening of monsoon and increasing climate aridity in the southern Loess Plateau during that period (Huang et al., 2000; Liu et al., 2011). Fire activity was in turn influenced by changing vegetation types, themselves influenced by seasonal precipitation variability at centennial-to-millennial scales.

Based on archaeological studies in the region, we propose that the gradual synchronous increasing trends in fire inferred from black carbon and charcoal records in the middle Holocene may be attributed to rapid climate variability (i.e. 6000–5000 years BP), rather that to increases in human activity during that interval. Since 3000 years ago, frequent fires appear to reflect local variations in effective moisture as well as regional heterogeneity in human land use and ignition sources.

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