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Environmental variability within the Chinese desert-loess transition zone over the last 20000 years

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Abstract: The desert/loess transition zone in northern China is sensitive to climate variability, which is controlled mainly by the relative strengths of the East Asian summer and winter monsoons. Sandy loess layers found in the Loess Plateau and palaeosol sequences found in the sandy desert demonstrate latitudinal shifts of the southern desert margin over the last 20000 years. Stratigraphic investigations together with radiocarbon and some thermoluminescence dating, show that during the last glacial maximum the desert margin was at its most southerly position (38°N). During the early Holocene, it moved northward about three degrees in latitude (41°N). At present, the desert margin is again close to its most southerly position (38°21′N), yet the northern boundary position of modern summer monsoon activity is placed at 41°45′N, which is close to the estimated Holocene Optimum desert margin (41°N). This situation cannot be explained from natural climate models. Hence, an external driving mechanism needs to be considered, and the most obvious one is that caused by human activity on the natural environment over the last 3000 years. Historical evidence tends to reinforce this consideration.

Key words: Human impact, natural environment, Loess Plateau, desert-loess boundary, monsoons, desertification, China.

Introduction

Messerli *et al.* (2000) have reviewed the impact of human activity on natural environmental fluctuations, and demonstrated that it has become a significant problem globally. China has a high population density, and hence the potential for the occurrence of human-induced environmental change. Detailed studies of past natural environments combined with present observations can provide a valuable framework for separation of human-induced changes from natural climatic cycles (Dodson and Intoh, 1999; Dodson *et al.*, 1993). The climate-sensitive northern boundary of the Loess Plateau in China provides us with a valuable resource for this type of exercise. The environment on the Chinese Loess Plateau is closely linked to inland Asian desert climate patterns. This semi-arid to arid area is contained within the 200–400 mm annual precipitation belt, and is sensitive to variations in East Asian monsoon circulation (Zhou *et al.*, 1996). It is dominated by dry grassland and shrubs, while forested areas are concentrated mainly along river corridors and deltas (An *et al.*, 1991a). Mobile, semi-mobile and fixed sand dunes characterize the arid landscape. During summer, a steady flow of moist maritime air from the southeast onto the continent produces about 80% of the annual precipitation (An *et al.*, 1991a). Hence, enhanced summer monsoon conditions over time encourage soil formation and vegetation development, resulting in sand-dune stability. During winter, a flow of cold, dry air out of north central Asia (winter monsoon) results in a significant flux of aeolian sand toward the southeast. The desert/loess boundary (or southern desert margin) is thus dominated by northerly winds generating frequent dust storms and deposition of aeolian sediments. Hence, sand, loess and palaeosol sequences from this area can be used as tracers to determine fluctuations in the position of the desert-loess boundary. One study

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of longitudinal shifts in the desert margin has been carried out (Sun *et al.*, 1998). This study has followed an east–west transect, covering a limited area of the actual loess-desert boundary, with a chronological framework that needs to be enhanced to cover the detail expressed in the stratigraphic sequences. The present polar and monsoon fronts are aligned in a northeast to southwest direction, as is the present southern desert margin. We focus on the positions of the desert margin for both the last glacial maximum and the Holocene precipitation optimum. Using both geological evidence and a detailed chronological framework obtained from a large number of ^{14}C and some TL age determinations, we have been able to evaluate desert margin position changes from a northeast–southwest transect of monsoon-sensitive sites.

Site descriptions and regional relationships

A series of 10 profiles along the desert-loess transition zone are shown in Figures 1 and 2. These were selected in order to examine past boundary conditions in detail. Five are representative of desert conditions. These are Cagelebulu (39°53′N, 108°18′E), Xijing (39°4′N, 105°8′E), Dunkou (40°33′N, 107°0′E), Sunitezuoqi (43°14′N, 113°52′E), and Xilinghaote (43°56′N, 116°3′E). During the last glacial maximum, these sites were under full desert conditions, and it is only the Holocene sediments (soils developed within aeolian sand) that indicate climate variation. The Holocene Optimum, in the early Holocene (~10000–5000 ^{14}C yr BP) for these sites can be bracketed between about 8000 and 5000 ^{14}C yr BP, as shown by the dashed lines in Figure 1.

The Cagelebulu Site has a depth of about 6.3 m (Figure 1), and the Holocene stratigraphy consists of sequences of aeolian sand, loess and palaeosol layers overlying lacustrine silt. Loess and soil development (more favourable conditions) occurred between ~9500 and 5000 ^{14}C yr BP, as indicated by the ^{14}C ages shown in Figure 1. These are based on the total soil organic matter fraction, after carbonate had been removed by acid leaching. The ^{14}C determinations were carried out within the Lanzhou Institute of Desert Research, Chinese Academy of Sciences (Dong *et al.*, 1995). The Xi Jing Holocene sediments have been developed on palaeosand dunes. The profile contains palaeosol material from 0 to 0.62 m (with the top of the profile being eroded away), then aeolian sand. The palaeosol ranges in age from 7870 ± 70 to 8090 ± 130 ^{14}C yr BP, based on the total organic fraction, after removal of carbonate. The ^{14}C ages were determined within the Lanzhou Institute of Desert Research, Chinese Academy of Sciences (Li, 1993). Dunkou is located 20 km north of the Xi Jing site, and the Holocene sediments consist of lacustrine mud intercalated with aeolian sand. The bottom of the profile (9.75 m) is gravel, overlain (9.75–5.8 m) by a lacustrine mud sequence containing shells. At the 6.5 m level, shell material has been dated at 10620 ± 100 ^{14}C yr BP, organic carbon at 7400 ± 110 (4.3–4.35 m) and 6220 ± 130 (2.25–2.30 m) ^{14}C yr BP respectively (Lanzhou Institute of Desert Research; Shi, 1991; Li, 1993). The Sunitezuoqi Holocene sediments consist of clay from 4.35 m to 2.8 m, overlain by aeolian sand from 2.8 to 1.65 m, palaeosol from 1.65 to 0.9 m, then aeolian sand. The palaeosol ranges in age from about 6300 to 7350 ^{14}C yr BP (age determinations on total soil organic matter, carried out within the Lanzhou Institute of Desert Research; Li, 1993). The Xilinghaote Holocene sediments consist of sandy palaeosols intercalated with aeolian sand containing relatively large quantities of calcite and gypsum. The palaeosols seemed to consist mainly of humic substances. Hence, pretreatment of palaeosol samples for the ^{14}C determinations involved successive leaching with 0.1 M NaOH solution, followed by leaching of the humic acids with a HCl/HF solution to remove any fine-grained clays closely associated with the humic acids. Finally, the fractions were converted from sodium humate to

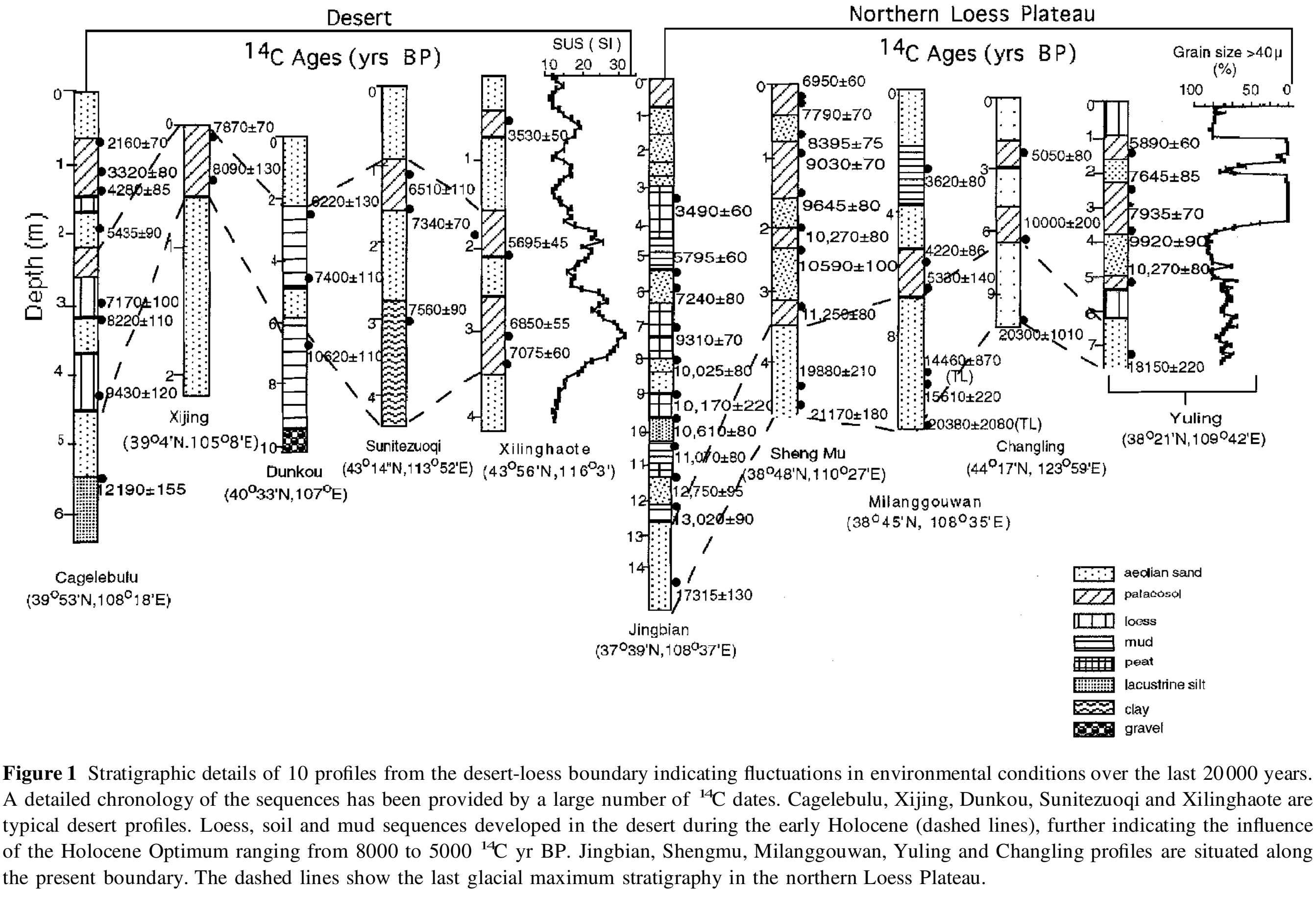

Figure 1 Stratigraphic details of 10 profiles from the desert-loess boundary indicating fluctuations in environmental conditions over the last 20000 years. A detailed chronology of the sequences has been provided by a large number of ^{14}C dates. Cagelebulu, Xijing, Dunkou, Sunitezuoqi and Xilinghaote are typical desert profiles. Loess, soil and mud sequences developed in the desert during the early Holocene (dashed lines), further indicating the influence of the Holocene Optimum ranging from 8000 to 5000 ^{14}C yr BP. Jingbian, Shengmu, Milanggouwan, Yuling and Changling profiles are situated along the present boundary. The dashed lines show the last glacial maximum stratigraphy in the northern Loess Plateau.

humic acid by cation exchange using Dowex 50 resin (Tan, 1996). The ^{14}C determinations were carried out using accelerator mass spectrometry by the NSF–Arizona AMS facility, University of Arizona. The results, as indicated in Figure 1, indicate two phases of soil formation, occurring at about 7200 to 6500 ^{14}C yr BP, and 5700 to 4500 ^{14}C yr BP. At this site, magnetic susceptibility (SUS) determinations were carried out at 2 cm intervals. SUS is an indicator of the relative summer monsoon intensity (An *et al.*, 1991b). It has been argued that the main carrier of the susceptibility signal is ultrafine-grained magnetite produced during *in situ* pedogenesis, which can also be linked to summer monsoon precipitation (Maher and Thompson, 1991). The magnetic susceptibility determinations verify the existence of the palaeosols, and also indicate the possible existence of a palaeosol at about 1.2 m (in the 4500 ^{14}C yr range).

The other five sites from the southern margin of the desert-loess transition zone (Figures 1 and 2) are representative of the northern Loess Plateau. In this case, during the last Glacial maximum, all sites represented in Figure 1 were within the desert boundary, with aeolian sand being deposited. Midiwan at Jingbian (37°39′N, 108°37′E), consists of 13.8 m of intercalated aeolian sand, silty mud, peat and palaeosol. A detailed description has been published previously (Zhou *et al.*, 1996; 1999). Wood, charcoal and peat samples were collected for ^{14}C dating. Wood cellulose was isolated from fossil wood samples using solvent extraction and sodium chlorite leaching as described by Head *et al.* (1989) and Zhou *et al.* (1990). Charcoal fragments were selected using a microscope and treated by acid, alkali and acid leaching (Donahue, 1993). The peat samples were sieved using an 80 μm sieve, and the fine (<80 μm) fraction was used for dating. This fraction consisted mainly of plant fragments, charcoal and pollen (Zhou *et al.*, 1996). Yangtaomao at Shengmu (38°48′N, 110°27′E) consists of intercalated sequences of aeolian sand and palaeosols (Figures 1 and 2). The site was described in detail by Zhou *et al.* (1996). Samples were collected for ^{14}C dating, and the fraction dated was either soil humin or the fine-grained organic fraction. Soil humin was prepared by solvent extraction and NaOH leaching (Head *et al.*, 1989; Zhou *et al.*, 1990), and concentrated pollen fractions were prepared using the method described elsewhere (Zhou *et al.*, 1997). Milanggouwan (38°45′N, 108°35′E), along the Ordos Plateau, contains aeolian sand, palaeosol and lacustrine mud sequences. ^{14}C determinations were carried out on the total organic fraction. Details of these and the thermoluminescence ages are given by Li (1993). Changling (44°17′N, 123°59′E), is much further east than the other sites, consisting of aeolian sand with two palaeosol sequences. ^{14}C dates are from the total organic fraction (Li, 1993). The Yuling site (38°21′N, 109°42′E), is between Jingbian and Shengmu, and is 5 km north of Yuling city. The profile consists of aeolian sand, loess and palaeosol sequences. The ^{14}C ages were carried out on the pollen fraction of each sample, and pollen (plus fine-grained plant material) separation was carried out as for the Jingbian

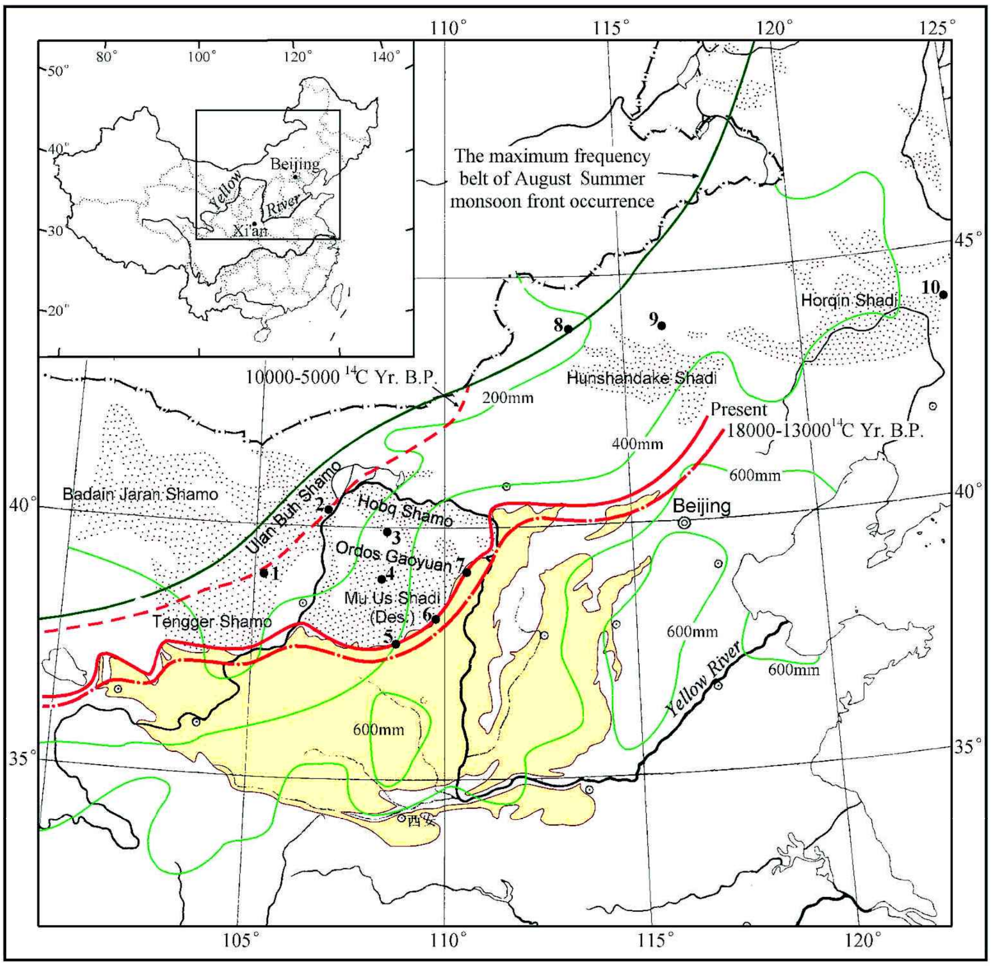

Figure 2 Reconstruction of the desert-loess boundary positions. The inset map indicates the main study area. The yellow shaded area indicates the Loess Plateau, while deserts are indicated by red stipple. Boundary positions (red) during the last glacial maximum (~38°, dashed line) and the early Holocene (~41°N, dashed line), and their relationship with the present day boundary (38°21′N, solid line) are indicated. Numbers 1 to 10 indicate the study locations: (1) Xijing (39°4′N, 105°8′E); (2) Dunkou (40°33′N, 107°0′E); (3) Cagelebulu (39°53′N, 108°18′E); (4) Milanggouwan (38°45′N, 108°35′E); (5) Jingbian (37°39′N, 108°37′E); (6) Yuling (38°21′N, 109°42′E); (7) Shengmu (38°48′N, 110°27′E); (8) Sunitezuoqi (43°14′N, 113°52′E); (9) Xilinghaote (43°56′N, 116°3′E); (10) Changling (44°17′N, 123°59′E). Present mean annual precipitation is shown in green.

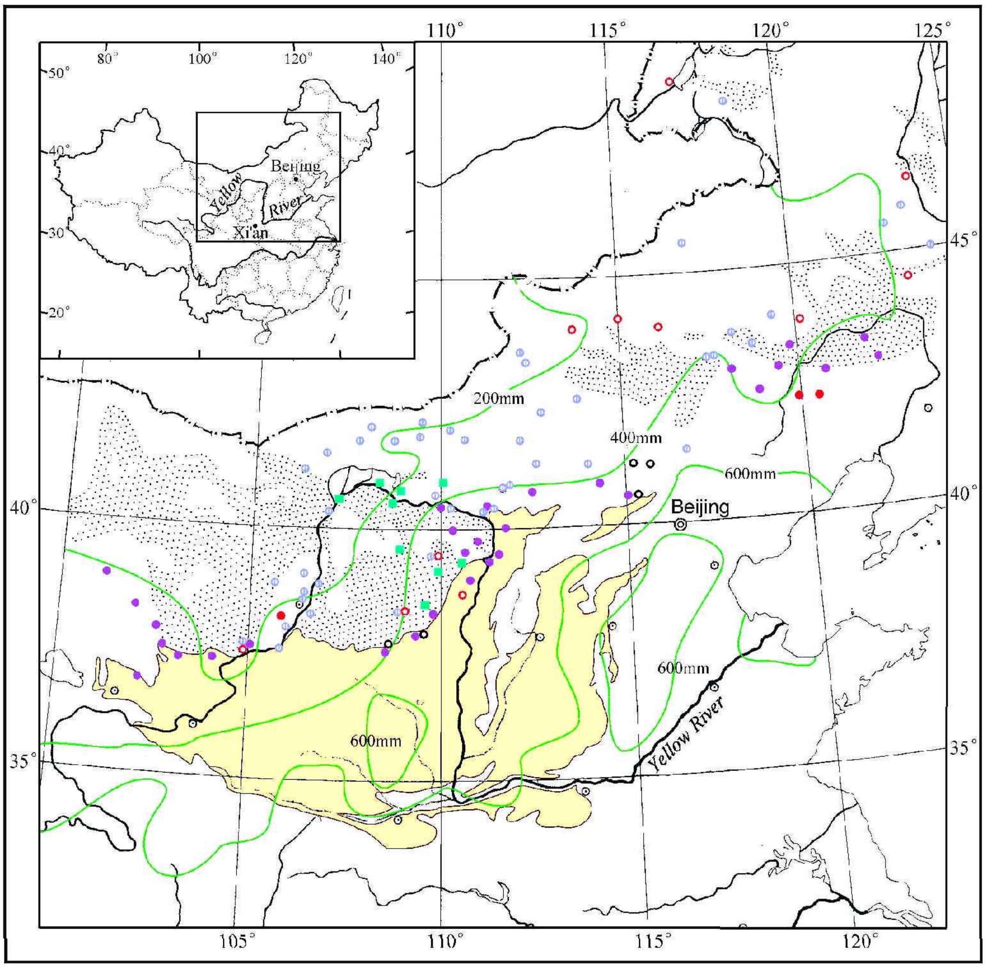

Figure 3 The distribution of prehistoric cultures along the Yellow River area. Microlithic culture (split blue circles, 9000–4000 ^{14}C yr BP), Yangshao culture (full red circle, ~7000–4000 ^{14}C yr BP), other cultures (opened red circle, ~4000 ^{14}C yr BP), recent habitation locations (blue squares, since ~3000 ^{14}C yr BP). The density of human occupation sites suggests that the environment was not desert, but probably grassland at that time. Other details are as in Figure 2.

profile. The method was detailed by Zhou *et al.* (1997). The ^{14}C determinations were carried out at the NSF–Arizona AMS Facility at the University of Arizona. The Yuling site was also sampled at 5 cm intervals for grain-size analysis, and the results are shown in Figure 1. Aeolian sand >40 μm found in the desert-loess transition zone (northern Loess Plateau) has not been affected by pedogenic processes, and is generally well sorted (Zhou *et al.*, 1996). It has previously been shown that the presence of a grain size component larger than 40 μm can be used as a winter monsoon proxy (An and Porter, 1997), and hence as a proxy of desert advance. The grain-size analysis from Yuling (Figure 1) illustrates the overall pattern for the five northern Loess Plateau sites. During the last glacial maximum (20000 to 13000 ^{14}C yr BP), aeolian sand (80% >40 μm) deposition occurred, indicating the extent of the southward shift of the desert margin under severe winter monsoon conditions.

The desert margin shifts for the two climate extremes have been reconstructed for the northern Loess Plateau and the southern desert margin (Figure 2). During the LGM, the desert margin shifted southward to a latitude of about 38°N (calculations based on the Ordos Plateau), not far from the present-day desert boundary (38°2′N, solid line). At the same time, the sea level along eastern China was 150–160 m lower than at present and the coastline was extended eastward some 800–1000 km (Wang, 1995). The Hobq Shamo (desert) and Mu Us Shadi (sandy land) were combined and the area of desert increased significantly. During the Holocene Optimum, the desert margin shifted about three degrees in latitude northward (~41°N; Figure 2) on the Ordos Plateau. ^{14}C AMS determinations on soil organics, loess and pollen from the desert sites ranged in value from ~5000 to 9600 ^{14}C yr BP, and magnetic susceptibility values were high. These higher values signify the presence of strengthened summer monsoon activity (Zhou *et al.*, 1998), except for one brief cooling period (Alley *et al.*, 1997). At this time, the extent of ice-affected areas was reduced, and sea levels were higher than present (Wang, 1995). The Mu Us Shadi was characterized by grassland. During the late Holocene (5000 ^{14}C yr BP to present), cooler conditions with some warm-wet fluctuations occurred (Zhou *et al.*, 1994; 1998).

Influence of human activity on the desert margin

In monsoon China, precipitation is related to the position of the summer monsoon front which at present is about 41°45′N in latitude, as shown in Figure 2 (Gao, 1962). The latitudinal positions of the desert margin postulated for the periods from 10000 to 5000 and 18000 to 13000 ^{14}C yr BP are about 41°N and 38°N respectively. The present desert margin is not far from that postulated for the LGM, yet the present monsoon front is in a similar position to that of the Holocene Optimum. This discrepancy allows us to test the hypothesis that human impact has influenced the position of the desert margin considerably, by considering the archaeological record.

We can assume that the early-Holocene climate (10000–5000

^{14}C yr BP) in China was dominated by natural variability (Lin, 1985). We can then postulate that the southward shift of the desert margin by about three degrees of latitude to the present is a result of the impact of anthropogenic activity, which includes inappropriate land use. Figure 3 shows the distribution of ancient cultural sites from about 10000 ^{14}C yr BP. The microlithic culture (two semi-circles, ~9000–4000 ^{14}C yr BP), the Yangshao culture (black circles, 7000–5000 ^{14}C yr BP); and other cultures (empty circles, ~4000 ^{14}C yr BP) were extensive within the area. More recent sites (squares, from 3000 ^{14}C yr BP) are closely aligned to the Yellow River. Most of the older excavated sites are now covered in sand. The relative density of these sites indicates that the environment was suitable for human occupation, though they seem to be clustered into two groups. One group was concentrated in the northern part of the semi-arid transitional zone, and is characterized by a microlithic culture consisting of fishing, hunting and animal husbandry. The other is characterized by dry farming agriculture, and was concentrated along the rivers and lakes. During the Holocene Optimum, highlands were selected for habitation, suggesting increased precipitation and river levels. At around 7000–5000 ^{14}C yr BP, during the Yangshao culture period, advanced coloured pottery making occurred in conjunction with primitive agriculture and domestic livestock management until about 4000 ^{14}C yr BP (Cao, 1986; An, 1987). From that time, this culture would not have survived if the environmental conditions were similar to those of today. At around 3000–2000 ^{14}C yr BP, the natural vegetation on the landscape was severely affected by an increase in human population (Cao, 1986), dry farming activity, forest clearance and frequent warfare (Wang, 1985). Wind and water erosion became a problem, and desertification became intense. The grassland ecology in some places was changed into semi-desert or desert, and mobile sand dunes developed.

Zhu *et al.* (1986) has detailed historical evidence of human activity within what is now the Mu Us Desert (Figure 2). At the beginning of the fifth century AD, Tongwan, the centre of the west Xia Dynasty, was built in the northern area of Jingbian County, within what is now the Mu Us Desert. Historical information indicates that the natural environment surrounding Tongwan was fertile land with plenty of available fresh water. From the establishment of the Tang Dynasty (AD 618), garrison troops and peasants cleared the natural undergrowth and planted grain in the southern area of the Mu Us Desert, south of the Ancient Great Wall. By the middle and late Tang Dynasty, because of frequent warfare causing much destruction, agricultural land was abandoned. The farmland and abandoned irrigation ditches became covered with sand transported by winter monsoon winds. Existing forests were burned during warfare, and natural grassland was trodden by war-horses. By AD 822, sand dunes were as high as the town wall of Tongwan on strong windy days. The change in the environment that had occurred during the lifetime of the town was extremely rapid.

From this period onwards, alternation between livestock and grain farming caused the opening-up of more land for farming, more wood required for fuel, and more livestock for grazing. By the middle of the sixteenth century, the Great Wall had been buried by sand in places, and could no longer be used for defence. Further land reclamation during the Ming and Qing dynasties, together with alternation between agriculture and animal husbandry, and winter monsoon winds played havoc with the natural landscape processes, resulting in the complete desertification of the Mu Us area. The time range of the desertification process can be documented from the late Tang (AD 705–907) Dynasty, through the Song (AD 960–1280) and Ming (AD 1368–1644) Dynasties, to the Qing (AD 1661–1911) Dynasty.

The present situation

It has been estimated that the area of desertification within North China has increased to approximately 117960 km^2, occupying 57.0% of the total area (semi-arid-arid transition zone) since historical time, and especially during the last 300 years, based on LANDSAT-TM images and calculations (Zhu and Wang, 1992). The Mu Us Shadi (desert) area has increased to 9400 km^2 in the last 50 years, covering 24.8% of the total area (Wu and Ci, 1998). This is occurring within a zone receiving between 200 and 400 mm rainfall per year. There is a clear picture of human activity as an external force, taking part in the natural environment change process. This picture is significantly clarified by the comparison that has been made with LGM and Holocene climate reproduction, though no comparison can be made with sedimentary sequences not affected by human activity over the last 3000 years, as very few remain in existence. The destruction of the environment caused by human activity is obviously a serious problem in this area. Health problems (fluorine poisoning, Keshan's disease, etc.) caused by either an excess or a scarcity of essential trace elements in the soils, because of poor land use, have been extensively documented (Liu *et al.*, 1985; Fordyce *et al.*, 2000). Hence the influence of human activity must be taken into consideration when predictions of future environmental changes along this sensitive ecological boundary are made. No doubt future research will find a two-way feedback existing between climate and anthropogenic impacts, and future ecological planning will need to take this into account.

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