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Holocene climatic and environmental changes in the arid and semi-arid areas of China: a review

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A HOLOCENE RESEARCH REVIEW

Abstract: This paper reviews recently published literature, most of which was published in Chinese, and searches for regional patterns of Holocene changes useful in depicting global patterns. The Holocene in the Xinjiang region can be divided into three stages: a warming and dry early stage (from 11 ~ 10 to 8 ~ 7 ka BP), a warm and wet middle stage (from 8 ~ 7 to 4.5 ~ 3 ka BP) and a fluctuating cool and dry late stage (since 4.5 ~ 3 ka BP). The Holocene in the northern Tibetan Plateau can also be divided into three stages: a warming and wet stage (from 10.5 ~ 10 to 5 ~ 4 ka BP), followed by a variable drying and probably warm stage (5 ~ 4 to 3 ka BP) and ending with a cool and dry stage (since 3 ka BP). In the Inner Mongolian Plateau, the early Holocene (from 10.5 ~ 9.5 to 8 ~ 7.5 ka BP) was warming and dry, and a warm and wet climate occurred from 7.5 to 3.5, during which the best time was 6.3–3.8 ka BP; the climate has been variably drying and probably cooling since 3.5 ka BP. In the northwestern part of the Loess Plateau, several Holocene palaeosols have been identified (10–9, 7.5–5, 4–3 and 2.7–2 ka BP) with the 7.5–5 ka BP palaeosol being most strongly expressed. The best-developed palaeosol-equivalent in major valleys is a swamp-wetland facies deposited between 8885 and 3805 ^{14}C yr BP under an extremely wet regime. The climate has fluctuated significantly at least three times around a dry and probably cool regime after the swamp-wetland facies-depositional period. Our summary shows that the Holocene Climatic Optimum occurred nearly contemporaneously (8–5 ka BP) at all sites in the Xinjiang region, in the Inner Mongolian Plateau and in the northwestern part of the Loess Plateau. A warming and wet early Holocene (10–8 ka BP) in the northern Tibetan Plateau is most likely related to high effective soil moisture resulting from snow and ice melting. We propose here that the middle Holocene Climatic Optimum (8–5 ka BP) in arid to semi-arid China was primarily a delayed response of the low latitude oceans to high latitude peak insolation (9–8 ka BP).

Key words: Holocene climates, arid environment, semi-arid enviroments, monsoon, Climatic Optimum, western China.

Introduction

The climatic history of the Holocene is of great scientific interest in understanding the future of the warming world. For example, to assess the present and future human-induced climatic changes that are superimposed on the natural trend, we have to first understand the natural trend during the Holocene (Overpeck, 1993; Rind and Overpeck, 1993). Until the trend is reliably portrayed, the human impact superimposed on the natural trend cannot be adequately defined (eg, Kutzbach and Liu, 1997; Bradley, 2000; Crowley, 2000; Delworth and Knutson, 2000; Levitus *et al.*, 2000; Hoerling *et al.*, 2001). Yet, our knowledge regarding the most recent geological past is quite limited. The most trusted Greenland ice-core record of Holocene temperature revealed a relatively stable Holocene climate (GRIP Members, 1995). But, the geochemical impurity data from the ice core show that the Holocene climate has been unstable, at least over the Greenland area (O'Brien *et al.*, 1995; Mayewski *et al.*, 1997). Holocene climatic instability has been reported from many places (Bianchi and McCave, 1999; Enzel *et al.*, 1999; von Grafenstein *et al.*, 1999; Wurster and Patterson, 2001; Baker

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et al., 2001; Luckge *et al.*, 2001; McDermott *et al.*, 2001) and different mechanisms controlling or modulating the Holocene climatic changes have been proposed (Bond *et al.*, 1997, 2001; Shindell *et al.*, 2001; Rind, 2002; Visbeck, 2002). To depict the temporal and spatial patterns of the Holocene climatic changes and associated environmental changes, we need to reconstruct the temporal patterns of different regions, especially those of climatically sensitive regions (Broecker, 1997).

The arid and semi-arid east-central Asia that includes China has been demonstrated to be one of the most sensitive areas to large-scale climatic changes (Li *et al.*, 1988; Thompson *et al.*, 1989; Feng *et al.*, 1993, 2005; D'Arrigo *et al.*, 2000; Jacoby *et al.*, 2000; Feng, 2001). It is also most likely to contribute to large-scale climatic and environmental changes through altering albedo-modulated or/and aerosol-modulated energy balance on land (Overpeck, 1993) and probably also through altering aerosol-stimulated organic fixation in oceans (Falkowski, 1997; Feng *et al.*, 1999). Our knowledge of Holocene environmental changes in arid and semi-arid China is far from satisfactory for depicting the temporal and spatial patterns in relation to global Holocene changes. However, persistent efforts during the past 20 years, especially during the past 10 years, have greatly improved our understanding of Holocene climatic changes. This paper aims at summarizing and scrutinizing the information, most of which was published in Chinese, and searching for the regional patterns of the Holocene changes in the hope that they can be useful in depicting global patterns. Arid and semi-arid China refers to the northwestern portion of China whose southeastern boundary is delineated by the 1.5 isoline of aridity (Figure 1).

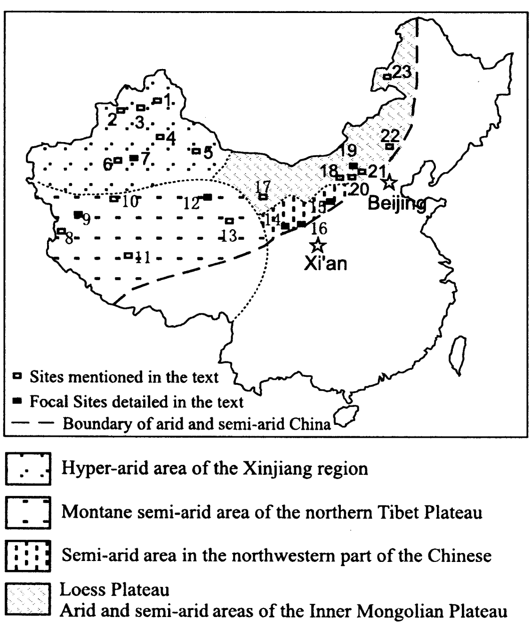

Figure 1 Map of arid and semi-arid China showing four physiographic regions and sites mentioned in the text. 1, Wulungu Lake; 2, Aibi Lake; 3, Manas Lake; 4, Mosuowan; 5, Balikun Lake; 6, Tarim River; 7, Bosten Lake; 8, Bangong Lake; 9, Sumxi Lake; 10, Guliya Ice Core; 11, Selin Lake; 12, Dunde Ice Core; 13, Qinghai Lake; 14, Lanzhou; 15, Yulin; 16, Dingxi Site; 17, Shiyang River; 18, Chasuqi; 19, Diaojiaohaizi; 20, Dahai Lake; 21, Huangqihai Lake; 22, Hunshadak; 23, Hulun Lake

Spatial patterns

To quantitatively reconstruct Holocene climatic and environmental changes, choosing appropriate proxies and establishing reliable chronologies are the keys to success. Nevertheless, neither the proxies nor the chronologies are yet sufficiently adequate. Most of the widely used proxies were adopted from western literature and were either not regionally tested on their bioclimatic dependencies or the bioclimatic dependencies were either not convincingly established. A good example is the well-studied summer monsoon proxy, magnetic susceptibility, in the Chinese Loess Plateau. The susceptibility signature has been described as a semi-quantitative proxy for summer monsoon intensity (An *et al.*, 1991) and even as a quantitative proxy for mean annual precipitation (Maher *et al.*, 1994). However, the relationships between past climates and the susceptibility signature are far from straightforward, especially at the level of temporal resolution required to identify Holocene climatic variability. For example, the magnetic susceptibility highs could be caused by either *in situ* pedogenic enhancement (Zhou *et al.*, 1990) or detritus inheritance (Feng, 1996; Sun and Liu, 2000). Dilution by carbonate concentration (Heller and Liu, 1986), magnetic enhancement by leaching processes (Anderson and Hallet, 1995), magnetic alterations by redox cycles (Feng *et al.*, 1998; Feng and Chen, 1999) and even particle-size redistribution (Han and Jiang, 1999; Feng *et al.*, 2004a,b) may also contribute to the finalization of the susceptibility signature. In brief, even this, the best-studied climatic proxy, cannot be used as a quantitative or even semi-quantitative proxy to reconstruct past climatic history (Feng and Johnson, 1995; Feng, 1996; Feng and Chen, 1999; Feng *et al.*, 2004b), not mentioning many other proxies that have not been properly tested in arid and semi-arid China. In addition, pollen analysis deserves a special mention here since it is regarded as a reliable, although qualitative, environmental proxy. Several attempts have been made to develop transfer functions expressing the relationship between the modern pollen rain and associated vegetation (Tang and Shen, 1996; Sun *et al.*, 1996; Liu *et al.*, 1999). However, the development of transfer functions is limited in China owing to the fact that intensive and long-term anthropogenic disturbances in the Inner Mongolian Plateau and the Chinese Loess Plateau have almost completely altered the natural vegetation (Zhao *et al.*, 2004). Furthermore, in arid basins surrounded by mountains (eg, in Xinjiang Province and the Hexi Corridor of Gansu), the transfer functions may not be applicable owing to the fact that not all of the Holocene pollen grains often recovered from fluvial or/and lacustrine sequences are produced *in situ* (Zhu *et al.*, 2001). Transfer functions are more or less reliable in the northern Tibetan Plateau as a result of somewhat less anthropogenic disturbances and lack of altitudinal vegetation zonation there (Tang and Shen, 1996).

Another key problem in Holocene climate reconstruction is the reliability of chronology. Two factors are responsible for the larger technical and inherited errors involved in C-14 dating. First, the best C-14 dating targets (eg, charcoal) are normally hard to find in the Holocene sequences of arid and semi-arid China. Second, commonly used C-14 dating targets, such as bulk organic matter, are inherently unreliable. For example, downward penetration of plant roots can mix time-diachronous organic matter and insect and animal activities can bring older organic matter upward (Birkeland, 1999).

With a full consideration of the climatic proxy resolution and chronological validity, this review attempts to summarize and scrutinize the published literature and is expected to add regional information for depicting the global Holocene climatic changes. To search for the spatial patterns of major climatic events in arid and semi-arid China, this paper reviews all work published during the past 10 years with foci on four physiographically different regions: (1) the hyper-arid area of the Xinjiang region; (2) the montane semi-arid area of the northern Tibetan Plateau; (3) the semi-arid area in the northwestern part of the Chinese Loess Plateau; and (4) the arid and semi-arid areas of the Inner Mongolian Plateau (Figure 1). To review the published literature on Holocene climatic changes in the four different physiographic regions, four factors are taken into consideration: (1) geographic coverage; (2) recent date (past 10 years) of publication in major scientific journals; (3) acceptability of the proxies used; and (4) likely validity of the chronologies established. For those focal sequences that are discussed in greater detail, higher standards are applied in terms of acceptability of the proxies and reliability of the chronologies. That is, the justifications for the proxies used are scrutinized and the validity of the chronologies is further examined (eg, dating targets and associated strata).

Xinjiang region

The northernmost site studied in the Xinjiang region is Wulungu Lake (site 1 in Figure 1). Yang and Wang (1994) show that the postglacial warming had started by 12 ka BP and the early Holocene was warm and dry. The highest lake-level of the Holocene appeared approximately from 7 to 5 ka BP under a warm and wet climate. This was followed by a dry and cool period (5–3 ka BP). The climate fluctuated along a mild and dry trend since 3 ka BP. At Aibi Lake (site 2 in Figure 1) a comprehensive investigation of geochemical indices and pollen assemblages was conducted (Wu, 1995; Wu *et al.*, 1996). A high A/C (*Artemisia*/Chenopodiaceae) ratio and higher ratios of unstable over stable chemical elements suggest that the early Holocene (10–8 ka BP) was dry and probably warming. A warm and wet climate dominated the middle Holocene (8–3.5 ka BP) and a mild and dry climate characterized the late Holocene (since 3.5 ka BP). At Manas Lake in northern Xinjiang (site 3 in Figure 1) a desert-steppe landscape was replaced with a steppe landscape during the period between 9 and 4 ka BP, suggesting a warm and wet climate (Sun *et al.*, 1994). The steppe landscape then deteriorated to a desert-steppe landscape at about 4 ka BP. The geochemical data of Lin *et al.* (1996) suggest the following climate sequence in the Manas Lake area: 12–10 ka BP, warming and dry; 10–6 ka BP, warm and wet; 6–5 ka BP, warm and dry; 5–3.5 ka BP, warm and wet; and 3.5–0 ka BP, fluctuating around a dry and cool regime. A recent study (Chen, H. Z. *et al.*, 2001) on a sand/soil sequence at Masuowan section in northern Xinjiang (site 4 in Figure 1) shows that the climate was dry and cool from about 10 to 7.5 ka BP. The best-expressed soil was developed between 7.5 and 4.5 ka BP. The period from 4.5 to 3 ka BP was a transition from a warm-wet climate to a cool-dry climate and the climate fluctuated drastically three times around a cool and dry regime from 3 ka BP on.

At Balikun Lake site in southern Xinjiang (site 5 in Figure 1), the Holocene climate was reconstructed from a multiproxy investigation (Han and Qu, 1992), which included analyses of pollen assemblages, trace elements, $\delta^{18}O$ and ostracodes. Their study indicated the following climatic sequence: 10–8 ka BP, warm and dry; 8–4 ka BP, warm and wet during which two relatively cool spells occurred in 8–6 ka BP and in 5–4 ka BP. The climate since about 4 ka BP was interpreted to have fluctuated considerably. A stratigraphically convincing case was presented in the middle reach of the Tarim River in the Tarim Basin (Feng, Q. *et al.*, 1999). The stratigraphic correlations of 11 fluvial–aeolian sections (site 6 and its vicinity in Figure 1) and relatively well-controlled chronologies (conventional C-14 dates of well-isolated litter layers embedded in fluvial strata) made this study more appealing, although its climatic proxy resolution is naturally limited because of its fluvial origin. They concluded that the postglacial warming led to an increase in the river discharge from 12 to 10 ka BP because of increased snow and ice melting in the headwater area. The warming period was followed by a 2000-yr cool and dry period (10–8 ka BP) when sand dunes were reactivated. The stratigraphic evidence shows that frequent floods and high discharges in the Tarim River characterized the period from 8 to 3 ka BP. Yet, gypsum formation of the same period suggests a dry climate in the river valley. After a cool and wet period (3–2 ka BP), sand dunes were reactivated under a very dry climate in the middle reach of the Tarim River.

Many studies have been carried out on Bosten Lake, our focal site in the Xinjiang region (site 7 in Figure 1). The lacustrine sediments have been analyzed at 5-cm intervals throughout a 3 m section and six conventional C-14 dates on well-isolated organic thin layers interbedded in sandy clay layers were obtained (Xu, 1998; Zhong and Xiong, 1998). The analysis of the laboratory data shows that $\delta^{18}O$ and $\delta^{13}C$ values of lacustrine carbonate are correlated in different ways. Wu and Wang (1997) argued, based on the relationships established in the eastern Tibetan Plateau, that $\delta^{18}O$ is a proxy of air temperature if the correlation is negative and that $\delta^{18}O$ is a proxy of lake level-related salinity if the correlation is positive, the latter being in good agreement with the carbonate content-indicated salinity. The negative correlation between $\delta^{18}O$ and $\delta^{13}C$ from 7 to 4 ka BP (Figure 2) suggests a warm middle Holocene (the shaded portion in Figure 2). A high A/C (*Artemisia*/Chenopodiaceae) ratio and a high C index ([Fe+Mn+Al]/[K+Ca+Mg]) suggest a relatively wet middle Holocene (7–4 ka BP). The climates in the early (10–7 ka BP) and late (4–0 ka BP) Holocene were characterized either by warm/dry combinations or by cool/wet combinations. The middle Holocene temperature–precipitation combination was distinctively different, eg, 7–5 ka BP was warm and wet and 5–4 ka BP was relatively cool and dry. To sum up, with a full consideration of the potential problems in proxy resolution and chronological validity, the Holocene in the Xinjiang region can be divided into three stages: a warming and dry early stage (from 11 ~ 10 to 8 ~ 7 ka BP), a warm and wet middle stage (from 8 ~ 7 to 4.5 ~ 3 ka BP), and a fluctuating but mainly cool and dry late stage (from 4.5 ~ 3 to 0 ka BP).

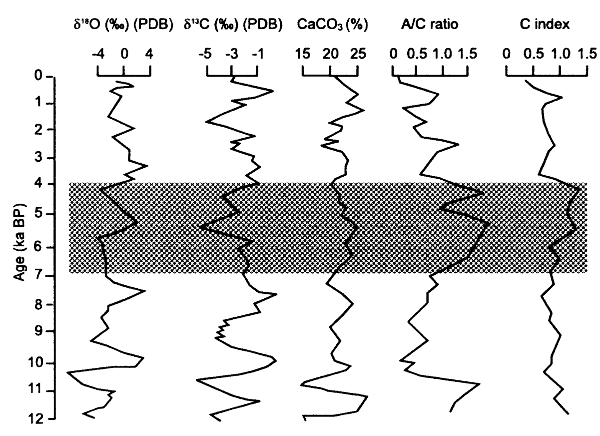

Figure 2 Variations in environmental proxies at Bosten Lake in the Xinjiang region. Note: $\delta^{18}O$ and $\delta^{13}C$ are the oxygen and carbon isotopic compositions of the lacustrine carbonate; A/C ratio = *Artemisia*/Chenopodiaceae; C index = $\Sigma(Fe+Mn+Al)/\Sigma(K+Ca+Mg)$

Northern Tibetan Plateau

Studies at Bangong Lake in the western margin of the northern Tibetan Plateau (site 8 in Figure 1) show that the postglacial lake-level rise was interrupted around 10.6–10.3 ka BP (Li *et al.*, 1991; Li *et al.*, 1994). The lake level then recovered to reach its recorded maximal level around 8.5–8.3 ka BP and maintained a high level until about 3.3 ka with a brief decline around 7 ka BP. Lake level has declined since 3.3 ka BP. At nearby Sumxi Lake (site 9 in Figure 1), a multiproxy investigation indicates that the lake level was high under a wet and probably warm climate from 10 to 4 ka BP, especially from 9 to 5 ka BP (Gasse *et al.*, 1991). A fall in lake level began around 5 ~ 4 ka BP and the climate entered the present situation (dry) around 3.4 ka BP, as demonstrated by mineralogical and geochemical indices and *Artemisia* percentage (a proxy for semi-aridness in arid areas) in the pollen spectrum (see Figure 3). The A/C (*Artemisia*/Chenopodiaceae) ratio suggests that the effective soil moisture was quite high from 10 to 6 ka BP (see Figure 3). $\delta^{18}O$ studies on the Guliya ice core (site 10 in Figure 1) indicate that the early Holocene temperature rose from 10.5 to 9 ka BP and then fell from 9 to 8 ka BP (Thompson *et al.*, 1997). The warmest period of the Holocene occurred around 7–6 ka BP and the temperature dropped significantly around 5–4 ka BP. Unlike records from other sites, $\delta^{18}O$-indicated temperature from the Guliya ice core started a general rise from 4 to about 1.5 ka BP (Thompson *et al.*, 1997; Yao *et al.*, 1997). At Selin Lake site (site 11 in Figure 1) the record shows an increasing trend both in temperature and precipitation during the early Holocene (10–8.4 ka BP). The warmest and wettest period of the Holocene occurred from 8.4 to 5.5 ka BP, which was followed by a moderate deterioration (cooling and drying) from 5.5 to 4.2 ka BP. The climate entered a drying stage from 4 ka BP (Gu *et al.*, 1993). Pollen data further suggest that the climate was generally wet from 9.6 to 3.8 ka BP with its climax around 7.5–6 ka BP (Sun *et al.*, 1993). A recent pollen investigation from the Dunde ice core (site 12 in Figure 1) demonstrates that the percentage of *Artemisia* was relatively high with a high A/C ratio from 10 to 4.8 ka BP (see Figure 3), indicating a more humid climate in the semi-arid northern Tibetan Plateau (Liu *et al.*, 1998). The relatively large variations in the *Artemisia* percentages from 4.8 to 1.5 ka BP indicate considerable climatic fluctuations. The $\delta^{18}O$-indicated temperature from the Dunde ice core was high from 7.5 to 1.5 ka BP (Yao and Shi, 1992). In the eastern margin of the Tibetan Plateau, a recent pollen study by Liu, X.Q. *et al.* (2002) from Qinghai Lake (site 13 in Figure 1) shows that the postglacial re-establishment of forests in the surrounding mountainous areas began as early as 10.5 ka BP, and both the tree pollen concentration and total pollen concentration reached a peak at 9.5 ka BP and a second peak at 7 ka BP with a significant vegetation deterioration around 8 ka BP. A drying trend started at 5 ka and reached its driest period around 3–2 ka BP (Figure 3).

To sum up, with a full consideration of the proxy resolution and chronological validity, the Holocene in the northern Tibetan Plateau can be divided into three stages: a warming and wet stage (from 10.5 ~ 10 to 5 ~ 4 ka BP), followed by a fluctuating dry and probably warm stage (5 ~ 4 to 3 ka BP) and ended with a cool and dry stage (from 3 ka BP on). It is notable from our discussion that, unlike in the Xinjiang region where the warm and wet climate occurred during the middle Holocene (from 8 ~ 7 to 4.5 ~ 3 ka BP), a warm (actually warming) and wet climate seemed to have occurred earlier in the northern Tibetan Plateau (from 10.5 ~ 10 to 5 ~ 4 ka BP), and it is in a good agreement with the record from the arctic Siberia (Kremenetski *et al.*, 1998; McDonald *et al.*, 2000). It should also be noted that the $\delta^{18}O$-indicated temperature trends from both the Guliya and Dunde ice cores exhibit three high-temperature periods (see Yao and Shi, 1992; Thompson *et al.*, 1997; Yao *et al.*, 1997), and the third high temperature period from 4 to 1.5 ka recorded in the ice cores has not been identified in the lake cores, perhaps because the proxies used lack the required sensitivity (Shi, 1997).

Northwestern Loess Plateau

The northwestern part of the Chinese Loess Plateau has a semi-arid climate (see Figure 1). Holocene loess/palaeosol sequences are selected here for comparison with the records from other regions: two sections (Landa and Jiuzhoutai) near Lanzhou (site 14 in Figure 1) and one section (Sandaoguo) near Yulin (site 15 in Figure 1). The loess/palaeosol strata and the corresponding variations in the magnetic susceptibility at the two sections near Lanzhou (Zhou *et al.*, 1991) document

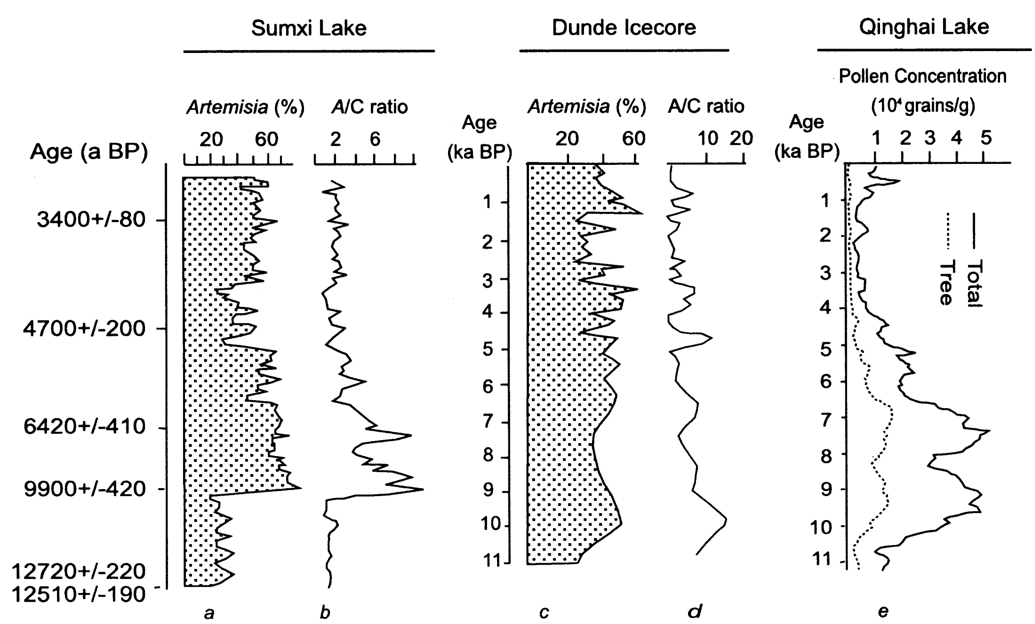

Figure 3 Variations in vegetation proxies in the northern Tibetan Plateau. Sumxi Lake core: *Artemisia* (%) and A/C ratio; Dunde ice core: *Artemisia* (%) and A/C ratio; Qinghai Lake core: tree pollen concentration and total pollen concentration

several soil-forming events: 10–9, 7.5–5, 4–3 and 2.7–2 ka BP (Figure 4A). Among them, the 7.5–5 ka BP event and the 2.7–2 ka BP event have the strongest magnetic susceptibility peaks. But, at the Sandaoguo section near Yulin (Gao and Chen, 1993; Gao *et al.*, 1996) three magnetic susceptibility peaks, formed approximately from about 5 ka BP to about 4 ka BP, contrast with a weakly expressed palaeosol bracketed by two dates (3.07 and 4.12 ka BP) at the Landa section near Lanzhou, and the minimally expressed palaeosol dated at 2.8 ka BP near Yulin contrasts with the well-expressed palaeosol dated at 2.38 ka BP at the Jiuzhoutai section near Lanzhou. Because of uncertainties in the chronologies (conventional C-14 dates of bulk organic matter) and qualitative nature of the magnetic susceptibility (eg, not comparable even

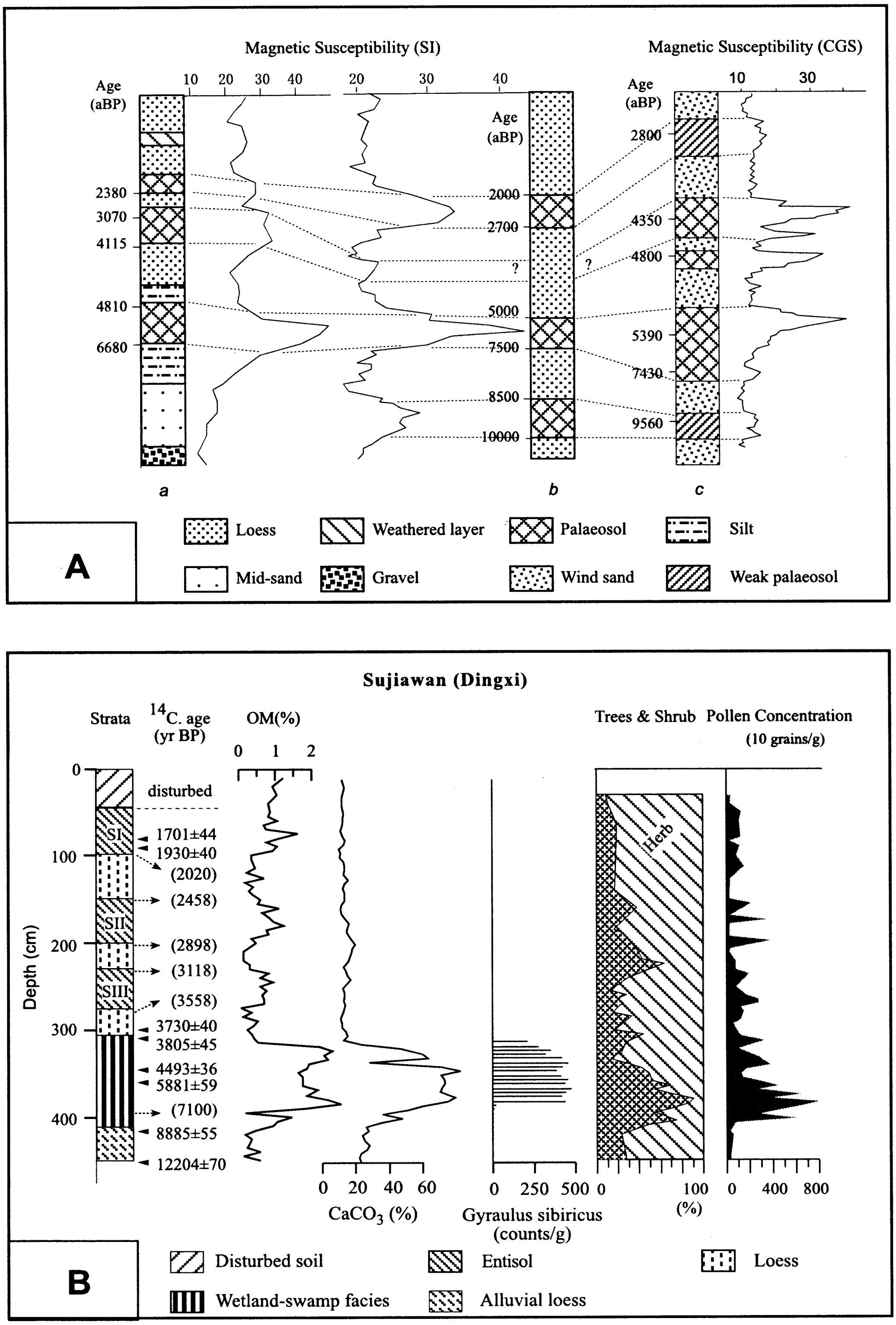

Figure 4 The Holocene sequences and proxy data in the northwestern part of the Chinese Loess Plateau. A. Loess–palaeosol sequences and the corresponding variations of magnetic susceptibility. a, Landa (Lanzhou); b, Jiuzhoutai (Lanzhou); c, Sandaogou (Yulin). B. Sujiawan section (Dingxi site) with: pedo-stratigraphy, AMS-dated ages and interpolated ages in parentheses. The proxy data include organic matter content (OM), carbonate concentration ($CaCO_3$), percentage of > 63 μm fraction (> 63 μm), aquatic snail *Gyraulus sibiricus* count, tree and shrub pollen versus herb pollen (%) and pollen concentration (10 gains/g)

between the two sections near Lanzhou), we are not able to provide more precise descriptions of Holocene climatic changes from these three sequences (Figure 4A).

A recent investigation on Sujiawan section (Dingxi site) in Gansu Province (site 16 in Figure 1) revealed much more detail of Holocene climatic changes in the northwestern part of the Chinese Loess Plateau (Feng *et al.*, 2004c). The Holocene sequence at the Sujiawan section (Dingxi) consists of two complexes: an upper loess–palaeosol complex and a lower wetland–swamp complex (Figure 4B). The lower wetland-swamp complex, which has been recorded as a ubiquitously occurring complex in the major valleys of the entire western part of the Chinese Loess Plateau and dated to have formed between 8885 and 3805 ^{14}C yr BP, is a greyish-blue clayey silt unit in which nearly horizontal sublayers were evidently affected by bioturbation. Aquatic molluscs (*Gyraulus sibiricus*) comprise much of the layer, about 25% volumetrically, and carbonate powders and half-decomposed organic matter (plant litter) are readily visible (Feng *et al.*, 2004c). The interpolated bracketing ages for three late Holocene palaeosols (ie, Entisols) within the upper loess–palaeosol complex are 3558 and 3118 ^{14}C yr BP for the Entisol SIII and 2898 and 2458 ^{14}C yr BP for the Entisol SII. The interpolated bottom age of the Entisol SI is 2020 ^{14}C yr BP (the interpolated ages are given in parentheses in Figure 4B). Pollen assemblages indicate the existence of a temperate forest during the early part (8885–4493 ^{14}C yr BP) and the existence of Compositae-dominating steppe during the later part (4493–3805 ^{14}C yr BP) with the interval between 7100 and 4493 ^{14}C yr BP being the best vegetated period of the Holocene (An, 2004; Feng *et al.*, 2004c, 2005). The subsequent loess unit (3805–3558 ^{14}C yr BP) and the overlying palaeosol SIII (3558–3118 ^{14}C yr BP) are palynologically characterized by steppe vegetation. A desert-steppe dominated the landscape from 3118 to 2898 ^{14}C yr BP. After a period of vegetation improvement (steppe) during the palaeosol SII formation (2898–2458 ^{14}C yr BP), a desert-steppe resumed (2458–2020 ^{14}C yr BP). A steppe landscape was again established around 2020 ^{14}C yr BP and appears to have deteriorated again around an inferred age of 1000 ^{14}C yr BP. To sum up, after the wet regime (8885 ~ 3805 ^{14}C yr BP), the climate has significantly fluctuated at least three times around a dry and probably cool regime.

Inner Mongolian Plateau

The Hexi Corridor in the eastern foothills of the northern Tibetan Plateau is included in the Inner Mongolian Plateau discussion here for the sake of convenience. The data from a well-studied aeolian–fluvial–lacustrine sequence (site 17 in Figure 1) in the Shiyang River Basin (Zhang *et al.*, 2000; Ma *et al.*, 2004) show that both temperature and precipitation started to increase around 10 ka BP and fluctuated between 10 and 7.5 ka BP. The climate was persistently warm and wet from 7.5 to 5 ka BP and more variably warm and wet from 5 to 3 ka BP. A massive reactivation of sand dunes after 3 ka BP suggests a drier climate. In the Inner Mongolian Plateau, the pollen data from a peat-silt section at Chasuqi (site 18 in Figure 1) show that the early Holocene (9.1–7.4 ka BP) was warming and relatively dry, and the middle Holocene (7.4–4.1 ka BP) was warm and wet. The late Holocene can be divided into four substages: (1) drying and cooling from 4.1 to 2.4 ka BP; (2) warming and increased wetness from 2.4 to 1.8 ka BP; (3) again drying and cooling from 1.8 to 1.4 ka BP; and (4) warming and increased wetness since 1.4 ka BP (Wang and Sun, 1997). A sediment sequence at Diaojiaohaizi (site 19 in Figure 1) has been studied by several researchers (Song *et al.*, 1996; Zhang *et al.*, 1997; Yang, 1998, 2001). Yang (2001) dated 13 separate peat layers (conventional C-14) and sampled the section at 1-cm intervals for pollen and geochemical analyses (Figure 5). Their data show that the early Holocene (10–9 ka BP) was a warming period. The period between 9 and 3.5 ka BP was warm and wet, as indicated by number of spore taxa, tree-pollen percentage and C index ([Fe+Mn+Al]/[K+Ca+Mg]). This warm and wet climate climaxed from 6.3 to 3.8 ka when the tree component was developed best (the shaded area in Figure 5). The late Holocene (from 3.5 ka BP on) was a drier and probably cooler period. Similar variations were also found in the nearby Daihai Lake core (site 20 in Figure 1). The data (Wang and Li, 1991; Li *et al.*, 1992) show that the lake level was generally high from 10 to 4 ka BP and the pollen-indicated Holocene Optimum occurred between ~ 8 and 4.4 ka BP (Xiao *et al.*, 2004; Xu *et al.*, 2004). The lake level has fallen since 4 ka BP with a brief rise between 2.4 and 2.1 ka BP (Wang and Wang, 1992). Another nearby sequence from Huangqihai Lake (site 21 in Figure 1) again exhibits similar variations: warming from 11 to 8 ka BP, warm and wet from 8 to 3 ka BP, and dry and cool since 3 ka BP (Li *et al.*, 1992).

A pollen analysis from Hunshandak Sandy Land (site 22 in Figure 1) shows that the climate was cool and humid in the early Holocene (10–8 ka BP), succeeded by a warm and wet climate from 8 to 5.9 ka BP. A warm and dry climate dominated from 5.9 to 2.9 ka BP and the climate then became drier and probably cooler since 2.9 ka BP (Liu, H.Y. *et al.*,

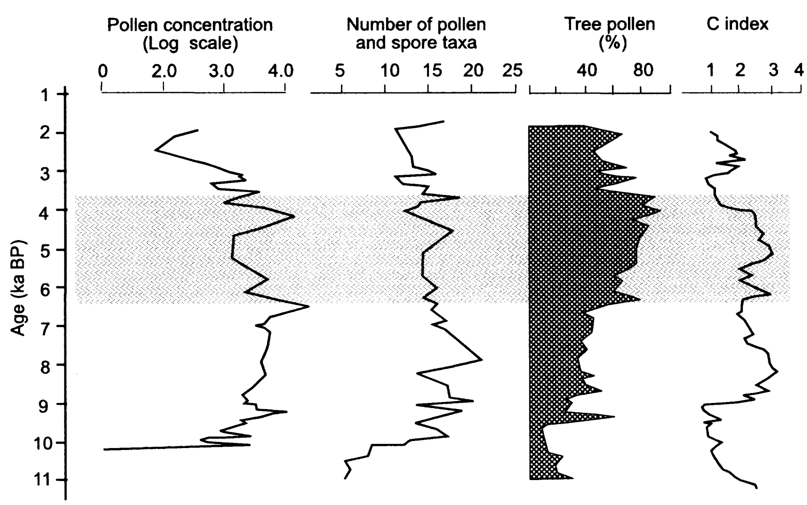

Figure 5 Variations in pollen concentration, number of tree and spore taxa, tree pollen percentage, and C index ($C = \Sigma(Fe+Mn+Al)/\Sigma(K+Ca+Mg)$) at Doaojiaohaizi site in the Inner Mongolian Plateau

2002). It should be mentioned here that a regional investigation of Holocene sand–dune mobility in the central part of the Inner Mongolian Plateau (Dong *et al.*, 1995) found that the sand dunes were relatively stable during the period between 7.5 and 5 ka BP and that late Holocene sand-dune reactivations were inferred to have occurred at 5–4.5, 3–2.5, 2–1.3, 0.8–0.5, 0.3–0.1 ka BP. At Hulun Lake (site 23 in Figure 1) in the northeastern part of the Inner Mongolian Plateau, a series of studies (Wang and Ji, 1995; Yang and Wang, 1996; Zhang and Wang, 2000) demonstrate that the early Holocene (10.3–7.2 ka BP) was a warming, yet dry period. A warm and wet climate (7.2–5 ka BP) and then a warm and dry climate (5–3 ka BP) characterized the middle Holocene. From about 3 ka BP on the climate became cooler and wetter. To sum up, the early Holocene (from 10.5 ~ 9.5 to 8 ~ 7.5 ka BP) in the Inner Mongolian Plateau was warming and dry, and a warm and relatively wet climate occurred from 7.5 to 3.5 ka BP, with 6–4 ka BP being the Holocene Climatic Optimum. The climate became variably drier and probably cooler since 3.5 ka BP at all sites mentioned except at the Hulun Lake site (site 23 in Figure 1) where the late Holocene climate (since about 3.0 ka BP) was cooler and wetter.

Discussion

Temporal comparisons of major events

The vegetation changes reconstructed at the well-studied Sumxi Lake site in the northern Tibetan Plateau show that the Sumxi Lake area (site 9 in Figure 1) was occupied by *Artemisia*-dominated semi-arid steppe vegetation during the early Holocene (10–6.5 ka BP) and deteriorated from 6.5 ka BP on, as indicated by the A/C ratio (Figure 6a). At another well-studied site, Qinghai Lake (site 13 in Figure 1), the tree pollen concentration, together with pollen assemblages (Liu, X.Q. *et al.*, 2002), indicates that the effective soil moisture was relatively high during the early Holocene (10–6 ka BP) with the period between 8 and 6 ka BP being the wettest (Figure 6b). At the Bosten Lake in the Xinjiang region (site 7 in Figure 1), a high A/C ratio, together with the negative correlation between the $\delta^{18}O$ and $\delta^{13}C$ values (Xu, 1998; Zhong and Xiong, 1998), suggests that the middle Holocene (7–4 ka BP) can be regarded as the Holocene Climatic Optimum (Figure 6c). It is also notable that frequent fluctuations in the A/C ratio at Bosten Lake site characterize the climatic variability since 4 ka BP. The pollen concentration at the Dingxi site (site 16 in Figure 1), as well as the pollen assemblage (Feng *et al.*, 2004c), indicates that the Holocene Climatic Optimum occurred from ~ 9 to 4 ka BP (Figure 6d) in the northwestern part of the Chinese Loess Plateau. At the Dajiaohaizi site in the Inner Mongolian Plateau (site 19 in Figure 1) the Climatic Optimum occurred in the period between 7 and 3.5 ka BP (Figure 6e; Yang, 2001).

However, a dry mid-Holocene (approximately from 7 to 4.5 ka BP) has been documented in two areas: (1) the northern tip of the Erdos Plateau represented by Yanhaizi Lake (around 41°N and 110°E) core data (Chen *et al.*, 2003) and (2) nearly the entire Alashan Plateau (37–42°N and 98–104°E) where dry mid-Holocene records were reported by Chen Fahu and his colleagues (Chen, F. H. *et al.*, 2001, 2004; Zhu *et al.*, 2001; Shi *et al.*, 2002) and more extensively by Herzschuh *et al.* (2004). Our reconciliatory explanation for the conflicting evidence against the extensively documented wet mid-Holocene is that the effective soil moisture during the mid-Holocene was probably indeed low in those extremely dry deserts where the increase in evaporation resulting from rising temperatures might have exceeded the overall increase in precipitation,

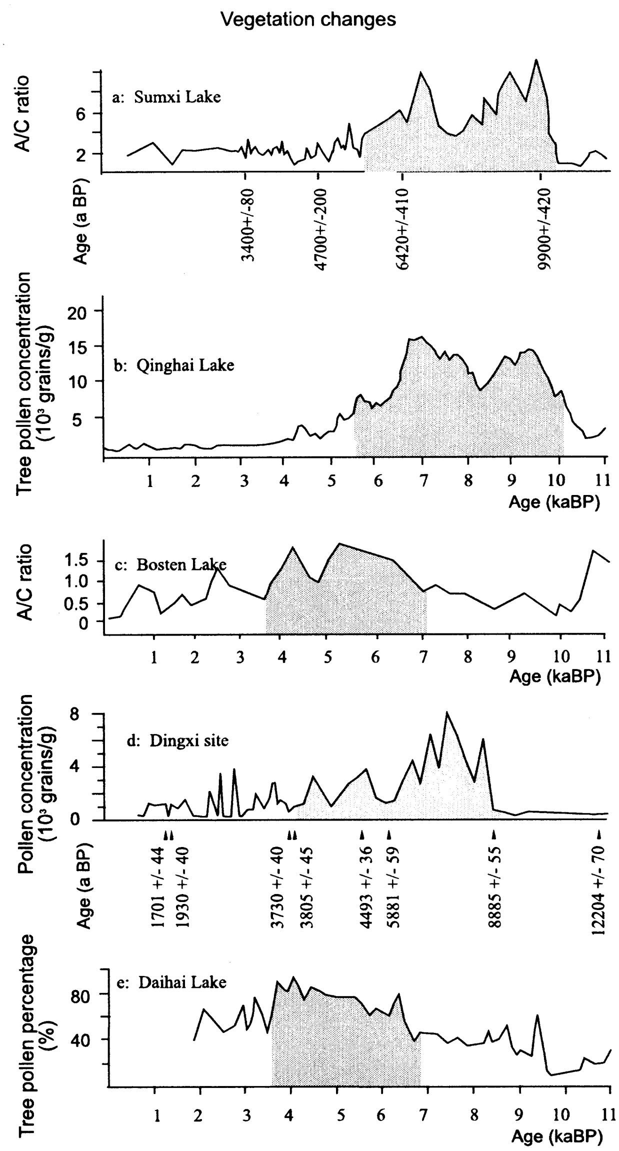

Figure 6 Temporal and spatial patterns of pollen-indicated environmental changes in arid and semi-arid China. a, A/C ratio at Sumxi Lake site in northern Tibetan Plateau (Gasse *et al.*, 1991); b, tree pollen concentration at Qinghai Lake site in the eastern Tibetan Plateau (Liu, X. Q. *et al.*, 2002); c, A/C ratio at Bosten Lake site in the Xinjiang region (Zhong and Xiong, 1998); d, total pollen percentage at Dingxi site in the northwestern part of the Chinese Loess Plateau (Feng *et al.*, 2004c); e, tree pollen percentage at Diaojiaohaizi site in the Inner Mongolian Plateau (Yang, 1998)

leading to an increase in climatic aridity and a decrease in effective soil moisture.

Mechanisms for producing the Holocene Climatic Optimum

Several recent studies highlight the middle Holocene (7–5 ka BP) as a period of particularly profound changes (see Steig, 1999). This also applies to arid and semi-arid China. The vegetation histories (Figure 6) indicate that the Holocene Climatic Optimum (warming and wet) occurred during the early Holocene in the northern Tibetan Plateau and the Optimum (warm and wet) occurred during the middle Holocene in the Xinjiang region, and in the northwestern part of the Chinese Loess Plateau and also in the Inner Mongolian Plateau. An and his colleagues (An *et al.*, 1993, 2000; Wu *et al.*, 1994) concluded that the variations in the Holocene East

Asian Monsoon were directly associated with the variations in the Northern Hemispheric insolation. The insolation peak at around 9–8 ka BP at 65°N in the Northern Hemisphere (Kutzbach and Gallimore, 1988) quickly initiated the East Asian Monsoon intensification (An *et al.*, 2000). Consequently, the East Asian Monsoon dominated the Inner Mongolian Plateau before 9 ka BP. The effect diminished toward the Yellow River Basin between 9 and 6 ka BP (see figure 13 of An *et al.*, 2000) and further toward the southeastern China between 6 and 3 ka BP (Figure 7A). Nevertheless, there are several flaws in the work of An *et al.* (2000). First, all dates of the East Asian Maximum (25 dates) northwest to the 6 ka BP line are not statistically different, especially considering uncertainties involved in bulk organic matter-based (conventional C-14) chronologies. That is, the age brackets (eg, 10–8, 10–8.5, 10–6, 9.7–8.5 ka BP) northwest to the 9 ka BP line are not significantly different from the age brackets (eg, 10–7, 10–8, 9.5–8, 8–7 ka BP) southeast to the 9 ka BP line. Second, the dates of the East Asian Monsoon Maximum southeast to the 6 ka BP line are even less convincing because even fewer data points were presented with even less-certain chronostratigraphies. Third, the Holocene Climatic Optimum of approximately the same timespan has also been reported from the Xinjiang region (see Zhong and Xiong, 1998), from the

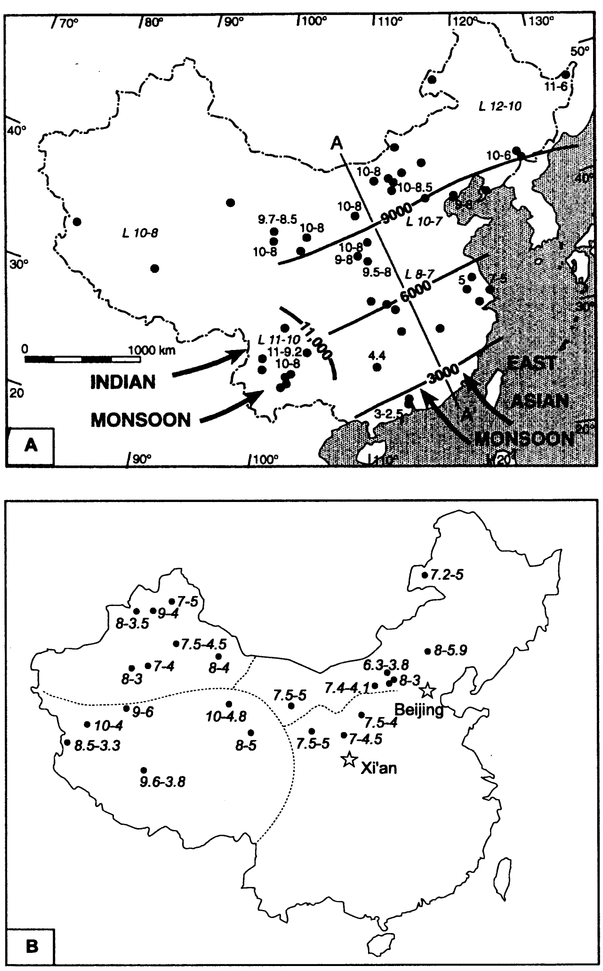

Figure 7 Chronological comparison of the reconstructed East Asian Monsoon Maximum (An *et al.*, 2000) and the reconstructed mid-Holocene Climatic Optimum (based on the literatures reviewed in this paper). A. Positions of the East Asian Monsoon Maximum through time based on palaeoclimatic proxy data (An *et al.*, 2000). B. Sketch map showing the reported dates of the Holocene Climatic Optimum in arid and semi-arid China (sites as in Figure 1)

northern Tibetan Plateau (see Yao and Shi, 1992), and even from the arctic Siberia (Kremenetski *et al.*, 1998; McDonald *et al.*, 2000) and from the northern Mongolian Plateau (Feng, 2001; Feng *et al.*, 2005), contesting the argument that the East Asian Monsoon Maximum had a time-regressive nature from the northwest to the southeast.

The East Asian Monsoon Maximum refers to the maximum precipitation or maximum effective precipitation (An *et al.*, 2000) and the Holocene Climatic Optimum refers to the Holocene warmest and wettest period (see Shi and Kong, 1992). Since the increased precipitation is the common item here, the temptation has been to equate the Holocene Climatic Optimum with the East Asian Monsoon Maximum. We plot the age brackets (or timespans) of the Holocene Climatic Optimum or the Holocene warm/wet period at 23 sites mentioned above to better portray the temporal and spatial patterns of the Holocene Climatic Optimum. The plot (Figure 7B) shows that the Optimum occurred nearly contemporaneously at all sites in the Xinjiang region, in the Inner Mongolian Plateau and in the northwestern part of the Loess Plateau, again with a full consideration of uncertainties in the chronologies. A warming and wet early Holocene distinguishes the northern Tibetan Plateau from the other areas. The A/C ratio at the Sumxi Lake site and at the Dunde ice core in the northern Tibetan Plateau shows that the effective soil moisture was indeed highest in the early Holocene. Our explanation for this exception is that the postglacial warming in this high-altitudinal plateau, as well as in high-latitudinal Siberia (see Kremenetski *et al.*, 1998; McDonald *et al.*, 2000), might be the factor that leads to increased soil effective moisture from melting snow and ice in the mountainous areas and probably also from melting permafrost in the valleys and basins. Thus, effective soil moisture cannot be equated with the effective precipitation in this case. Studies at Qinghai Lake in the northern Tibetan Plateau also caution against equating maximum precipitation with maximal effective precipitation (Liu, X.Q. *et al.*, 2002). For example, the first tree-pollen peak (around 9 ka BP, see Figure 6b) was probably caused by higher effective soil moisture under cool conditions (ie, from suppressed evaporation) as indicated by higher percentages of *Picea* and *Abies*. The second tree-pollen peak (around 7 ka BP, see Figure 6-b) was characterized by increased percentages of temperate forest components, implying high precipitation under warm conditions. Thus, both the inferred effective soil moisture and the inferred effective precipitation may be misused in reconstructing the timing of the East Asian Monsoon Maximum or the Holocene Climatic Optimum.

The early Holocene insolation peak at 65°N that occurred around 9–8 ka BP (Kutzbach and Gallimore, 1988) might have set the stage for the middle Holocene Climatic Optimum. That is, the peak insolation might not only have accelerated the postglacial ice sheet retreat in high latitudes and snow-cover disappearance in mid-latitude mountains but also rapidly warmed up the continental interiors. However, the 'oceanic buffering effect' (Clemens *et al.*, 1991, 1996; Feng *et al.*, 1998; Yao and Shi, 1992) might have played an important role in bringing in the water vapour from the low-latitude oceans to the mid-latitude lands, since oceans need much more energy to warm up than the land. It is thus likely that low-latitude oceans were gradually and sufficiently warmed from the peak insolation period (9–8 ka BP) to the Holocene Climatic Optimum (approximately from 8 to 5 ka BP) so that the enhanced evaporation effectively injected more water vapour into the East Asian Monsoon system and other precipitation-pro systems (eg, the westerlies). In addition, the increased amplitude of the seasonal cycle of insolation in the Northern

Hemisphere at about 6 ka BP might have resulted in an increase in the sea-surface temperature in the low-altitude oceans in late summer (Kutzbach and Liu, 1997). This increased sea-surface temperature around 6 ka BP might have also enhanced the East Asian Monsoon-related precipitation and made additional moisture available to other systems such as the westerlies. This scenario is directly supported by the maximum Holocene lake levels occurring in India (Enzel *et al.*, 1999) during the Northern Hemispheric increased summer insolation period (around 6 ka BP) and indirectly supported by the maximum Holocene aridity occurred in the southern America (Baker *et al.*, 2001) during the Southern Hemispheric decreased summer insolation period (8–5.5 ka BP).

Furthermore, the warmer and wetter climate resulted in denser vegetation cover during the Holocene Climatic Optimum, and the air–vegetation–soil coupling under the denser-vegetated wetter conditions might have further promoted the Optimum by feedback mechanisms, as convincingly suggested by observed data (eg, GLACE Team, 2004) and modelled data (eg, TEMPO, 1996; Doherty *et al.*, 2000; Chen *et al.*, 2002). Finally, the dry and warm middle Holocene (around 6 ka BP) in the mid-continental USA (Olson *et al.*, 1997; Dean and Schwalb, 2000; Mason and Kuzila, 2000) and the warm and wet middle Holocene in arid and semi-arid China (see Shi, 1992) make the long-term (millennial-to-centennial timescale) variations in ENSO intensity (Moy *et al.*, 2002) another possible scenario in explaining the Holocene Climatic Optimum. That is, the warm-water shift in a long-term La Nino-like system (or Super-ENSO as referred by Stott *et al.*, 2002) towards the Asian side of the Pacific might be able to generate a wet eastern Asia and a dry mid-continental USA, although the inferred high temperature at both sides of the Pacific needs further articulation before this scenario becomes a convincing one (Feng *et al.*, 2004c).

One more point we want to make is about the beginning and the ending of the mid-Holocene regime. The mid-Holocene is reportedly bracketed by 8.0–7.0 ka BP and 4.5–3.0 ka BP in the Xinjiang region, by 10.5–10.0 and 5.0–4.0 ka BP in the northern Tibetan Plateau, and by 8.0–7.5 and ~3.5 ka BP in the Inner Mongolian Plateau. These large ranges of the beginning and end may reflect the spatial variations in these vast physiographic regions, but they most likely resulted from the problems in the validity of the chronologies that we are not able to resolve here. We want to place our focus here on the aforementioned western part of the Chinese Loess Plateau where a sequence at Dingxi site (site 16 in Figure 1) was well dated. Specifically, the climatically wet and ecologically well-vegetated regime was bracketed by loess deposition at the beginning (~8885 ^{14}C yr BP) and at the end (~3805 ^{14}C yr BP). These sharp transitions are marked not only by field-observed stratigraphic contrasts (wetland/swamp layer versus loess) but also by laboratory analysed proxy data (eg, organic matter content, carbonate concentration, aquatic snail count and pollen assemblages). It is a surprise to see the timing consistency (Figure 8) between the climatically wet regime in the western part of the Chinese Loess Plateau (Feng *et al.*, 2004c, 2005) and the climatically warm regime in the arctic Siberia (between 60°N and 65°N). Kremenetski *et al.* (1998) and McDonald *et al.* (2000) reported that the postglacial boreal-forest development commenced by 10 000 cal. yr BP, advanced southward between 9000 and 7000 cal. yr BP and retreated northward between 4000 and 3000 cal. yr BP in arctic Siberia. The forest advancement was attributed to a number of temperature-related factors including heightened summer insolation, ice-sheet demise, sea-ice cover reduction and landward penetration of warm North Atlantic waters. The forest retreat resulted from declining summer insolation and cooling arctic waters under a neoglacial regime. The timing consistency between the western part of the Chinese Loess Plateau and the arctic Siberia implies that the East Asian monsoon-modulated precipitation in the western part of the Chinese Loess Plateau

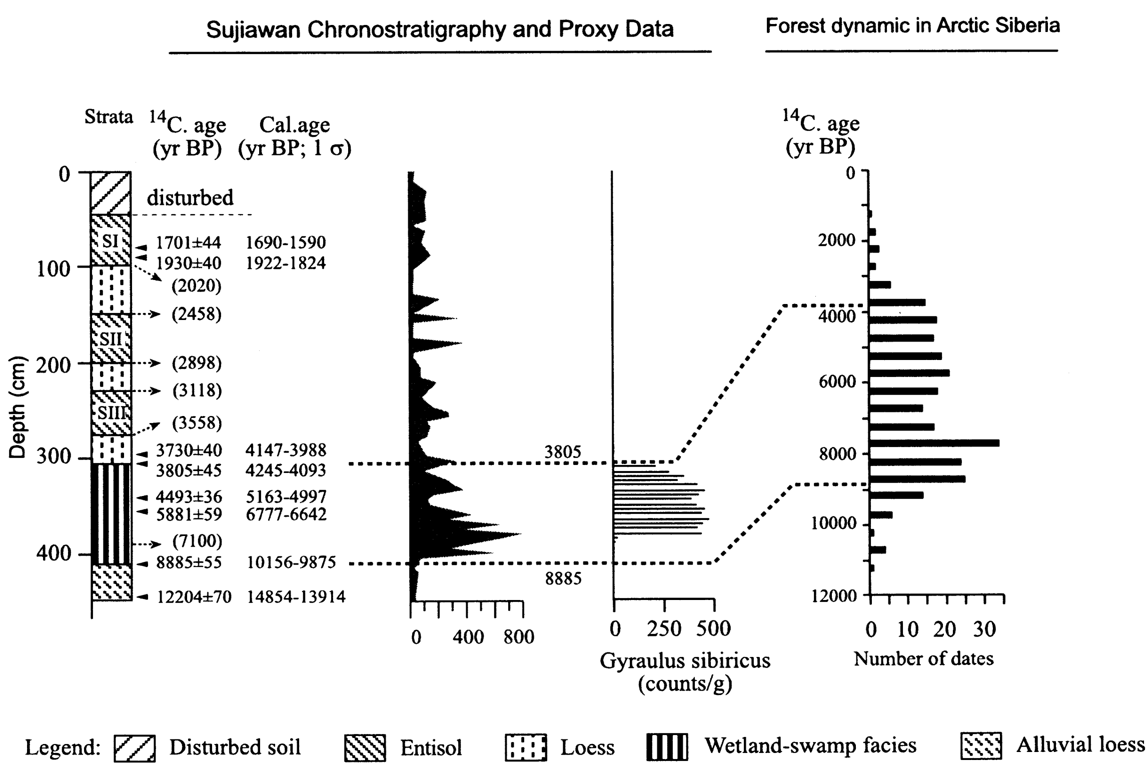

Figure 8 Proxy data at Sujiawan section of Dingxi site (site 16 in Figure 1) and the comparison with forest dynamics in the arctic Siberia (Kremenetski *et al.*, 1998). The proxy data include pedo-stratigraphy, AMS dates with calibrated dates, aquatic snail count (*Gyraulus sibiricus*), and pollen concentration (Feng *et al.*, 2004c)

has probably been causally related to high northern-latitude temperature.

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