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Paleoenvironments in China during the Last Glacial Maximum and the Holocene Optimum

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Two key-periods, the Last Glacial Maximum (LGM) and the Holocene Optimum represent respectively the coldest and warmest episodes that were testified in China by a great number of paleoenvironment evidences. At the LGM, the climate was colder and drier with the mean annual temperature 7–10 °C lower than today in the North and 4–6 °C in the South. The multidisciplinary data are consistent with global changes, such as the depression of snowline, southward advancing of permafrost limit, expansion of loess deposition, lowering of the lake level, invasion to the eastern China of desertic vegetation, displacement to the South of zonal vegetation, many discoveries in temperate climate zone of the northern cold mammals, and the remarkable sea regression resulting in the eastward shift of the Chinese coastal line. The data of Holocene recognized a warm period (ca. 9,000–6,000 yr BP) that the temperature estimates were generally 2–4 °C higher than nowadays. The rise of sea level resulted in great changes of the coastal environment in eastern China. Due to the warm climate, glaciers largely retreated and even disappeared in the eastern part. The permafrost was located to the North of the current limit. Desert and loess greatly reduced and the highest lake level was found mostly during 9,000–5,000 yr BP. The boreal conifers were to the north of the present locations, and many steppe areas were covered with forested formations. The above past climate and ecological pattern of the key-periods in China helps to better understand the future changes and their results.

Introduction

A good number of dated observations on the paleoenvironmental conditions in China during the Last Glacial Maximum (LGM) and the Holocene presently exist. However, most of the publications are in Chinese and not available to the greater part of occidental scientists, despite the efforts of a few authors (Shi, 1984; Liu, 1991; An et al., 1991b) who synthesised those data in some international journals. More recently, research by many international scientific groups provided more detailed paleoenvironmental records for some regions of China (Wake and Mayewski, 1996). The Chinese Quaternary paleodata were also synthesised in Winkler and Wang (1993). Here we shall only discuss the consequences of the last two extreme

global scenarios (the coldest and the warmest) on the diverse regions of China.

The principal paleoenvironmental features are represented: coastline, mountain glaciers, permafrost, aeolian formations, surface water, vegetation and climate. It is evident that such documents may only be a rough estimate of paleo-situations as inferred from literature : the purpose of this work is therefore to give a preliminary view of what is presently known of environmental changes in China during two key-periods.

Present-day environments in China

China has a total land area of 9.6 million km² and measures some 5,500 km from north to south. The Chinese mainland is flanked southwards by the Bohai, the Huanghai (Yellow Sea), the East China Sea and the South China Sea. The coastline of the mainland is about 18,000 km long. Since the Yellow and Yangze Rivers run into the west Pacific, forming deltas and coastal plains in the northern part of China, the geomorphology of the coastline divides into two sections : muddy coasts with wetland and lagoons to the north of the Hangzhou Bay (30°15' N), and estuaries and sandy coasts with hills and mountains to its south (Figure 1). It is obvious that the northern section of the coastline is more sensitive to sea level changes than the southern one. Another important feature is the wide extent of the continental shelf along the eastern coast of China, which stretches out from the coastline over 500 km.

The present snow-line is the highest in the Tibetan Plateau where it is located between 5,200 m and 6,200 m (Wang, 1984; Shi, 1991). However, it lowers both northwards and eastwards. For example, it lies between 3,700 m and 4,200 m in Xinjiang region and only at 2,800–3,350 m in the Altay Mountains (Table 1). Modern glaciers on the mountains of western China cover a total area of 58,651 km², about half of the mountain glaciers in Central Asia (Huang, 1991). The most expansive glaciers are in the Kunlun and Himalayas mountain ranges (Table 1). The glaciers in China can be geomorphologically classified into valley, cirque and hanging glaciers. The observations on glacial termini from the 60s to the 70s showed that 55 % of them were retreating, while only 26 % were advancing (Shi, 1984). More recent works in the Tianshan, Altay and Pamir show that about 95% of the glaciers had a negative mass balance over the past 30 years (Dyurgerov, 1994).

Continuous permafrost covers 2,150,000 km², i.e. nearly 22% of the Chinese territory. High latitude permafrost, located from 51°N northwards, is mostly found on the Daxinganling and Xiaoxinganling mountains of northeastern China (Wang, 1984). In the north of the Tibetan Plateau, the lowest limit of permafrost is at about 4,200 m (Wang and French, 1995).

Sand and gravel deserts are the major landforms in northwestern China. The total area covered by "gobi-deserts" is about 1,282,000 km², extending from 75°E to 135°E and from 35°N to 50°N, accounting for 14% of the total Chinese territory (Zhu, 1980). Among the largest deserts are the Taklimakan, Gurbantunggut, Badain Jaran, Tengger, Ulan Buh and Mu Us. About 8% of the

Table 1 Snow-line and glaciers in western China (after Huang, 1991).

Mountain range	Snow-line (m)	Number of glaciers	Glacier area (km ²)
Altay	2,800–3,350	416	293.20
Tianshan	3,600–4,300	8,908	9,196.98
Qilian Shan	4,400–5,250	2,856	1,973.50
Pamir	4,200–5,900	2,112	2,993.85
Kunlun Mts.	4,500–6,000	1,848	12,482.17
Karakorum	5,000–5,600	7,774	2,642.20
Qingtang Plateau	5,600–6,200	1,821	3,108.81
Gangdise	5,800–6,000	936	1,668.00
Nyainqentanglha	4,600–5,600	3,099	7,536.75
Tanggula	5,400–5,700	2,966	2,082.00
Himalayas	4,300–6,200		11,055.00
Hengduan Mts.	4,600–5,500	1,680	1,618.62

desertified land were formed during historical times. Mu Us desert, for example, was a vast grassland in historical times, and a city was built there in 413 AD. Wind-blown sand began to invade its suburbs by 828 AD and the city was eventually destroyed by aeolian deposition in 994 AD (Dong, et al., 1984; IGPD, 1980). Desertification is also well recorded in the Kolsin area, Inner Mongolia, where tall trees and luxuriant grasses were present until the mid-10th century. Nowadays, about 10% of its surface are covered with shifting sand (Hou, et al., 1982).

Loess is also a kind of remarkable landform in North-central China. The primary sources of loess are wind-born dust from desert areas. The Chinese loess plateau lies today between 30°N and 40°N and covers an area of about 300,000 km² with a thickness varying from 100 to 200 m (Liu, 1991). The thickest loess deposits are found in the western part of the Loess Plateau. The loess in Gansu and Shanxi Provinces covers the entire land surface and three types of landforms can be distinguished: the Yuan (high plain), the Liang (flat-topped range) and the Mao (dome-like hillock) (Wang, 1984; Liu, 1985).

There are about 50,000 rivers in China with a drainage area of more than 100 km². The Yangtze River, for example, covers a drainage surface of 1.8 million km², with a length of 6,300 km (Liu, 1984). Natural lakes are also common, and most of them are scattered in the middle-lower Yangtze River plains and on the Qinghai-Tibet Plateau. About 61% of the glaciers are located in regions of inland drainage, which results into numerous large and shallow lakes and closed basins surrounding the arid deserts in northwestern China (Huang, 1991). The total area of the marshlands covers 110 million hectares, and most of them are distributed in northeastern China and in the western part of the Sichuan Province. The rivers in the plains of northeastern China, for example, are bordered with wide paludal areas.

The most outstanding climatic features in China are the strong influence of the summer monsoon in the East, the arid and hyperarid conditions in the Northwest and the frozen environment on the Tibetan Plateau (Table 2). The Tibetan Plateau, appearing as a continent in an atmospheric ocean, imposes both dynamic and thermal effects on the climate, and influences the vegetal cover.

The SE monsoon circulation is driven by differential heating between the West Pacific Ocean and the Eurasian supercontinent. It is closely related with the precipitation repartition in China. Therefore, the climate

Table 2 Main climatic characteristics in China.

Regions		Mean annual temperature (°C)	Annual precipitation (mm)	Dominant wind direction	
Eastern monsoon region	Northeastern China	-5–7.7	500–800	S	N-NW
	Northern China	12.2–15.7	400–800	S	NW
	Southern China	16.3–23	1,000–2,500	S-SE	NW
Northwestern arid region		4–10	15–300	NW-N	S-SW
Tibetan Plateau		-4–11	50–600	W	W

in the East is humid or subhumid, with forest as the dominant vegetation (except for the strong human impact regions). The vegetation in the eastern monsoon zone of China belongs to temperate and sub-tropical zones. The natural forest formations, from north to south, are mainly cold conifer forest, conifer and deciduous mixed forest, deciduous broad-leaved forest, mixed broad-leaved forest, evergreen broad-leaved forest, evergreen monsoon forest and rain forest (Wu, et al., 1980).

The arid areas in northwestern China belong to the eastern section of the immense Eurasian deserts and steppes (Wang, 1981). However, some conifer (*Picea*) forests can be found in the Altay and Tianshan Mountains. As for the Tibetan Plateau, there is an increasing desiccation tendency from the southeast to the northwest, in close relation with the mountain block of moisture-bearing air masses, coming from the Indian Ocean and the Pacific. The vegetation there, is well forested on the south and southeastern slopes, but steppic on the high plateau and desertic in the Kunlun Mountains (Cui, 1988).

The Last Glacial Maximum

During the Last Glacial Maximum, sea level was 120 to 130 m lower than today, and the coastline of the East China had shifted by 600–1,000 km eastwards from its present location (Wang and Wang, 1980; An, et al., 1991). The Korean Peninsula, Taiwan and Hainan

Figure 1 Location of places quoted in the text.

Islands were connected to the mainland, resulting into the disappearance of Bohai and Huanghai (Yellow Sea). The South China Sea was much smaller and almost closed (Wang et al., 1995). The increased continentality then reinforced the drier and colder climate over the Chinese mainland.

According to the paleodata and the radiocarbon ages, it can be ascertained that the LGM snowline was lowered by 300–1,000 m on the mountains of western China. The snow-line depression was rela-

Table 3 Lowering of the permafrost limit in China during the Last Glacial Maximum (after Zhou, et al., 1991).

Mountain range	Latitude	Present-day permafrost limit (altitude)	LGM permafrost limit (altitude)
Altay	47–49 °N	2,200–2,800 m	1,000–1,600 m
Tianshan	44–45 °N	2,700–3,100 m	1,900–2,300 m
Qilian	37–40 °N	3,500–3,900 m	2,700–3,100 m
West Kunlun	36 °N	4,400–4,500 m	3,900–4,000 m
Central Kunlun	35 °N	4,150–4,200 m	3,600–3,700 m
East Kunlun	34.5 °N	3,700–3,800 m	3,200–3,300 m
Tangula	32.5 °N	4,600–4,650 m	4,100–4,200 m
Gongga	29.5 °N	4,600–4,900 m	3,600–3,900 m
Himalayas	28 °N	5,100–5,300 m	4,800–5,000 m
Wutai	38 °N	2,300 m	1,800 m
Taibai	34 °N	3,000 m	2,500 m
Changbai	42 °N	1,800 m	800 m

tively moderate in the inner regions of the Tibetan Plateau and of the western Kunlun mountains (300–500 m lower), whereas it was much greater on the southeastern flanks of the high plateau, particularly on the southern slopes of the Himalayas, reaching 1,000 m or even more (Shi, 1991). There is no doubt that glaciers widely expanded during the LGM in western China and, to a less extent, in eastern China, such as in the Taibai Mts (3767 m, 33°55'N, 102°21'E), Luoji Mts (4359 m, 27°N, 102°21'E), Xiao Hinggan (4045 m, 28°4'N, 102°E) and Changbai Mts (2749 m, 42°N, 128°E), and even like the Yushan Mts in Taiwan Island (3950 m, 23°30'N, 121°E). However, it is now generally agreed that an inland ice cover on the Tibetan Plateau did not exist during this glacial phase. Furthermore, a study of the glacial history of the central Himalayas shows that the glacial maximum there (ca 63,000 yr BP) does not coincide with the LGM of the northern hemisphere ice sheet (Sharma and Owen, 1996). Recently, several ice cores drilled on the Qinghai-Tibetan Plateau have been studied (Thompson et al., 1989; 1997). The relatively high anion and microparticles at the LGM from the Dunde ice core primarily reflect the dust influx from the Taklimakan and Qaidam deserts. Recent results also show that the $\delta^{18}\text{O}$ values decrease to their lowest levels during the LGM, which is consistent with the patterns in polar records (Thompson et al., 1997).

The permafrost limit in northeastern China evidently moved southwards during the LGM. Its southern limit extended from the southern Liaoning Province to the regions around 41–42°N, i.e. Liaodong Peninsula, Beijing, Taiyuan, Lanzhou and Hengduan Mts (Guo and Li, 1981). This assumption is supported by paleo-permafrost relics, including pingoscars, sand wedges, ice-wedge casts and polygonal soils or folded layers (Zhou, et al., 1991). The LGM permafrost southern limit matches the present temperature isotherm of 7–8 °C, suggesting a c 8 °C cooling in these areas. Table 3 indicates the lowest permafrost limit in the mountains, at present and during the LGM, evaluated from relic records and taking the snow-line depression into account.

Due to the strong winter monsoon, the desert areas in northwestern China were more extensive than at present, with the limits of

active sand dunes shifting eastwards and southwards. The sand dunes even invaded the steppe and sandy regions like Mu Us and Kolshin, as well as the lower ground of the Yellow and Yangtze rivers. The loess deposition became much wider, covering the middle and lower reaches of the Yangtze River and the mountain slopes to the west of the Sichuan basin. On the Chinese Loess Plateau, the loess deposited, during the LGM changes, from silty loess to clayey loess in a downwind (southeasterly) direction, as it becomes thinner. The well-developed loess and paleosols provide an excellent long-term climatic proxy record. Since the 1980s, the loess has been an interesting topic of research being one of the most imposing landforms in China which could be very sensitive to the winter monsoon variations during the glacial-interglacial intervals. A number of exposed loess sections in the central loess plateau has thus been studied by Liu (1985; 1991), Liu et al. (1986), Kukla and An (1989) and Porter and An (1995). The techniques used to study the evolution of the paleoenvironment include soil development, magnetic measurements, grain-size analysis, organic carbon content, mollusca shells, rare earth elements, microfabric and micromorphology (Wake and Mayewski, 1996). The loess sequence L1, known as the Malan loess, was deposited during the last glacial cycle under drier and colder conditions, and the substage L1-1 represents the LGM event (Liu, et al., 1986). The loess was deposited during the LGM (10–30 m) at an inferred average rate of 0.35 mm per year, which confirms the enhancement of dust accumulation (Dawson, 1992). The loess interdisciplinary studies also show that loess dust accumulation is more coherent with the variations of the global ice volume than with those of orbital insolation, suggesting that the aeolian dust deposition is causally related to the development of continental ice sheets (Shackleton et al., 1995; Ding et al., 1996).

The expansion of the loess deposits was coeval with the low level of inland lakes. A Late Pleistocene salt deposition was discovered in many sites such as the Baingoin lake in Tibet (Zheng et al., 1983), the Qarhan lake in the Qaidam basin, the Qagan lake in Inner Mongolia and the Aydingkol lake in Xinjiang (Li, et al., 1989). The study of a salt lake in the Qaidam basin (36–39°N, 91–99°E) with evidence of halite deposition and clastics identical in chemical composition to the Malan loess (Winkler and Wang, 1993), shows that the lake level was lower and the lake size remarkably reduced. The lakes Qinghai (Tibet) and Daihai (Inner Mongolia) also had low levels. Effectively, the LGM low lake levels were recorded not only in the northwestern regions of China, but also on the Tibetan Plateau (Van Campo and Gasse, 1993) and throughout eastern China, including the tropical regions (Fang, 1991; Zheng and Lei, in press). Many lakes in the middle and lower reaches of the Yangtze River reduced or even disappeared and were replaced by loess-like deposits (Wang and Li, 1991). However, some studies from Yunnan, in southwestern China, show that the lake levels were higher during the LGM in a number of basins like Kunming, Xinjie and Yuxi (NIGL, 1989).

A great number of pollen analysis indicate that both desert and steppe vegetation extended and invaded the eastern part of China during the LGM (Ke and Sun, 1992; Kong and Du, 1980; Kong, et al., 1994). A steppe vegetation in northeastern China is also indicated by the distribution of fossil mammals of *Mammuthus primigenius* (woolly mammoth) and *Coelodonta antiquitatis* (woolly rhinoceros) at more than 200 sites in the region (Qiu et al., 1981). The northern cold climate animals like *Cricetulus*, *Microtus brandti* and *Ursus arctos* spreaded to Yuhang in the Zhejiang Province and to Lishui in the Jiangsu Province, regions located to the south of the Yangtze River (Zheng and Han, 1991). In the temperate zone of eastern China, vegetation boundaries shifted deeply southwards. The boreal conifer (mainly *Picea*, *Abies*, *Larix* and *Betula*) forests replaced the temperate deciduous forest in a large part of eastern China (Winkler and Wang, 1993), and the tropical rain forest moved out of China (Zheng, 1991). Numerous researches focused on the Tibetan plateau demonstrate that a desert-steppe vegetation (mainly with *Chenopodiaceae* and *Ephedra* in the northwest, and *Artemisia* and *Cyperaceae* in the center and southeast) covered the greatest part of the Tibetan plateau (Tang and Shen, 1996; Van Campo and Gasse,

1993), and that the forest formation only existed in the extreme south and east of the plateau (Tang, 1996). The tree line was about 1,200 m lower than nowadays.

According to the paleoclimatic proxy data, the greater part of China was drier and colder. The annual mean temperature in northern and central China was at least 7 °C to 10 °C lower than today, as inferred from both pollen and permafrost limits (Zhou, et al., 1991; Sun and Chen, 1991), although some estimates were larger (10–12 °C lower) (Cui and Xie, 1985; An, et al., 1991). The lowering of the mean annual temperature in eastern Tibet was estimated to be 4.8 °C to 7.9 °C, but 6 °C for January in southeastern Tibet, where a 250 mm annual rainfall was inferred (Shen, et al., 1996). Similarly, in southern subtropical China, it was about 4 °C to 6 °C lower when mountain forest boundaries lowered (Zheng and Lei, in press). The annual precipitation in the north was 200 to 300 mm lower than at present (Kong and Du, 1981), and 400 to 600 mm lower in the regions presently under the monsoon influence.

Many records have proved that paleomonsoons varied at a large scale during the LGM (Yan and Petit-Maire, 1994). The loess sequence is one of the well-studied records containing paleomonsoon signals (An et al., 1991b). For instance, magnetic susceptibility and particle-size measurements of some soil samples from diverse localities of the Chinese loess plateau were used as indicators of variations in intensity of the East Asian winter monsoon (Ding et al., 1995). These works showed a clear increase in coarse-size loess grains that accumulated during the LGM, suggesting that the enhanced southeastern movements of cold polar frontal air produced a predominant spatial sorting of loess grains during a subaerial transport (Liu et al., 1989; Zhang, 1984). The strengthening of the winter monsoon is also obvious in eastern China and even in central Japan, where the aeolian quartz flux in lake deposits markedly increased at the LGM (Xiao et al., 1997). Some results also indicate that a stronger East Asian winter monsoon was coupled with cooler and drier conditions, in response to the increased continentality of the Chinese mainland, to sea level depression and to lower SSTs (Winkler and Wang, 1993; Wang and Wang, 1990; Zheng and Lei, in press).

The Holocene Optimum

The marine transgression rapidly flooded the continent since 11,000 yr BP, and reached its maximum at about 6,000–7,000 yr BP (Wang and Wang, 1980). Many deltaic sediment records, like coastal lacustrine and salt marshes, tidal flats and shelf deposits testify to strong marine and deltaic influences at the Holocene optimum, the absolute rise reaching about 2 to 3 m above the present-day sea level (Yang, 1991). The Holocene transgressive process deeply modified the coastline landscape to the north of Hangzhou Bay. The studies on Yangtze River deposits show that the transgression could affect areas upstream as far as 1,600 km from the present-day river mouth (Yang, 1991). Along the western coast of the Bohai Sea and of the Yellow Sea, complex deltaic deposits of the Yellow River were accumulated since the deglaciation and during the transgression: they cover nowadays a huge area of about 29,400 km² in northern China. This resulted into the displacement of the Yellow and Luanhe river mouths by 600 km at the most (Xue, et al., 1991). Therefore, it is evident that the changes of coastal environments were mainly controlled by a sea level rise prior to 6,000 yr BP, and by a delta progradation after 6,000 yr BP (Xue et al., 1995). In contrast, the coastline to the south of Hangzhou Bay was relatively stable during the Holocene optimum. The Zhujiang river (Pearl river) mouth became an estuary at 6,000–7,000 yr BP and, since then, was filled up with deltaic sediments (Xue, et al., 1991).

The glaciers in western China rapidly retreated, while those in eastern China completely disappeared, due to the warmer climate since the early Holocene. Three glacial fluctuation stages can be distinguished during the Holocene: a slow retreat at 10,000–8,000 yr BP, a rapid retreat at 8,000–3,000 yr BP, and alternating advances

and retreats from 3,000 BP on (Zheng and Shi, 1982). The warming also resulted in the degradation of permafrost and the northern shift of its boundary. However, permafrost could still be observed on the Tibetan plateau and in other regions of high altitude (Wang, et al., 1979). In northeastern China, the permafrost southern limit was located about 100 km to the north of its current situation. Studies of the Guliya ice core from Tibet show high $\delta^{18}\text{O}$ and CH_4 values throughout the Holocene, particularly in the early Holocene, which is correlative with warmer, moister conditions in the subtropics and argues strongly for a low-latitude methane source at that time (Thompson et al., 1997).

The desert and loess distributions were clearly reduced during most of the Holocene. A series of sections on the Chinese loess plateau indicate that the loess deposition rate lowered at the very beginning of the Holocene, and that the soils rapidly developed during the Holocene optimum, due to the less intensive winter monsoon circulation (Liu, 1985). In the sandy areas of Mu Us and Kolshin, dunes vegetated and consolidated and oasis were well developed during the Holocene megathermal (Gao et al., 1992).

Most lakes levels in China were high during the Holocene optimum. The highest levels in the northern and central regions were between 6,000 and 3,000 yr BP, and between 9,000 and 7,000 yr BP (Fang, 1991), therefore possibly recording a drier episode between 7000 and 6000 BP, as already observed by Petit-Maire and Guo (1996). In northeastern China and on the Tibetan plateau, many basins received ice-melt water to form larger lakes. The layers of freshwater mud found in Lake Zhacancaka (32°30'N, 82°20'E) indicate several millennial-scale wet periods, between 8000 and 5000 yr BP (Wang and Fan, 1987). Three Holocene lake terraces were also found above the present level of Lake Zharinamuco (31°N, 85°E) (Li et al., 1985; Winkler and Wang, 1993). Several lakes in western Tibet (Sumxi Co and Bangong Co) also reveal that lake levels were higher with warmer and wetter conditions during the Holocene megathermal (Gasse et al., 1991; 1996). The lake level records from Qinghai closed-basin show that the maximum level was at about 8,000 yr BP (Lister et al., 1991). Lakes Daihai and Huangqihai, in Inner Mongolia, show their highest levels between 8,500 and 6,000 yr BP with a warmer and wetter climate (Li et al., 1992).

A great number of palynological studies on the Holocene vegetation changes in China have been carried out since the 80s. In eastern China, a boreal conifer forest is believed to have moved north of the 50th parallel, i.e. about 200 km further north than nowadays. Temperate mixed and deciduous forests covered a large part of northern and northeastern China, i.e. about 5° further north than nowadays. Some sites from Hebei plain demonstrate an increase in deciduous broad-leaved forest elements between 9,000 and 2,000 yr BP (Yan, et al., 1981). The present-day steppe areas in Inner Mongolia were covered with a forest or forest-steppe vegetation (Li et al., 1992). The subtropical monsoon evergreen broad-leaved forest extended to the domain of the Yangtze River, at 35°–37°N (Sun and Chen, 1991). When sea level was higher at the Holocene megathermal, the tropical mangrove forest was very developed in the littoral areas of Fujian and Guangdong Provinces (Zheng, 1995). As for western China, a rise of arboreal pollen (mainly sub-alpine conifers) was observed in the Qinghai lake for the interval dated 8,000 to 3,500 yr BP (Du et al., 1989). On the central Tibetan plateau, a montane and alpine steppe with Cyperaceae, *Artemisia* and Gramineae, with some tree elements, widely developed between 9,600 and 6,000 yr BP. The *Artemisia* steppe or sub-desert occupied western Tibet, inferring wetter conditions and higher precipitation in comparison with the arid climate at the LGM (Tang and Shen, 1996). The continental fauna also demonstrates an optimal climate for the subtropical or Indo-malayan species such as *Hyaena*, *Acinonyx*, *Rhinoceros*, *Elephas* and *Pongo* (Zheng and Han, 1991). *Elephas maximus*, *Rhinoceros sondaicus* and *Tapirus indicus* lived as far north as 34°–41°N.

Temperature estimates for the Holocene optimum were 1 °C to 6 °C higher than today, depending on the geographical location. The general temperature relative to nowadays was, from roughly 8,500 to 5,000 yr BP, +2 °C to +4 °C after An et al. (1991), and +3 to +5 °C

after Zhang et al. (1997). Sun and Chen (1991) summarised that it might be 4 °C to 6 °C higher in some areas of Tibet, 3 °C to 4 °C in northern China, 2 °C to 4 °C in eastern China (Xu, 1989) and only 1 to 2 °C in southern China (Walker, 1986). A similar synthesis was also made by Shi et al. (1992), suggesting that the average annual temperature was about 4 °C–5 °C, i. e. 3 °C, 2 °C and 1 °C higher than today, respectively, on the Qinghai-Tibet Plateau cold temperate areas, warm temperate areas and subtropical areas. A recent study, using the pollen analogue concept to reconstruct paleoclimate, shows a synchronous change of positive anomalies in the whole of eastern China at 7,000–6,000 yr BP (Yu and Qin, 1997), the calibrated annual temperature being then 3.1, 1.8, 2.2 and 3.6 °C higher than today in the Northeast, the North, the Centre and the South of China respectively. If the above temperature estimates are exact, northern China should be one of the warmest regions in the northern hemisphere during the Holocene Optimum, if compared with the results calibrated by Velichko et al. (1991). The warmer climate in China and the higher SSTs in the Pacific and Indian Oceans strengthened the summer monsoon circulation, with more abundant, but contrasted seasonal precipitation (Yan and Petit-Maire, 1994). The annual rainfall in the southeastern monsoon areas was at least 100 mm higher than at present (An et al., 1991), and could reach 400 mm in some regions. The quantitative estimates in annual precipitation for the mid-Holocene, based on 113 pollen sites, were consistently greater than today (50–300 mm/yr) throughout eastern China (Yu and Qin, 1997). On the Tibetan plateau, the southwestern monsoon began to increase at 11,000 yr BP, and reached its maximum at around 7,000 yr BP (Tang and Shen, 1996). The pollen ratio of *Artemisia/Chenopodiaceae*, a moisture indicator, was the highest between 7,700 and 4,300 yr BP in western Tibet (Van Campo, et al., 1993).

The future

A global climatic change is most probably current, due to the rapid increase of radiative gases (CO₂, CH₄, and N₂O) in the atmosphere. In addition, chlorofluorocarbons (CFCs), which have heat-holding capacities 15,000 times greater than CO₂, are also being added by man to the troposphere. The enhancing of the natural greenhouse effect by this pollution should, according to models, induce a global warming of +1 °C to +3 °C in the 21st century. As has been proposed for the Sahara desert (Petit-Maire et al., 1990), one can assume that the effect of this man-made rise of global temperature will probably be of the same nature (even if not of the same degree) as past climate warmings, despite the fact that its causes are different and not an analogue by all means. An activation of the monsoons would be the first consequence of any warming events (Petit-Maire, 1990). The climatic and environmental future would then develop towards a scenario related to the one observed for the Holocene optimum, which corresponds to a global warming of about 2 °C and to Chinese estimates of +2 °C to +4 °C relative to nowadays. The future climatic changes will certainly be much more abrupt and extended than in the early Holocene, since the last step of deglaciation and the effects of orbital forcing upon the earth continental environments were generally quite slow at human scale, while very rapid at geological time. The following oncoming problems should lead us to pay attention:

1. Abrupt climatic changes induce a rupture of the present-day stable situation and bring out, before a new stability is reached, severe natural hazards such as extra-tropical cyclones, floods, droughts, land-slides, etc. Such events were already observed at least one decade ago (Flohn, 1987, 1989).

2. During the early Holocene, human population was scarce and, except in coastal areas and river plains, little concerned with slow climatic change. Instead, the highly populated modern China is liable to be severely affected by the rapid oncoming environmental changes, particularly in the flat deltaic regions like the northern coastal areas of China where even a moderate sea-level rise (estimated to some 0.5 m in 2050) would have severe effects.

3. Moreover, it is difficult to transpose past *natural* changes of *natural* landscapes into future *man-made* changes of *man-made* landscapes, as is the case in China.

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