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# Timing and spatial distribution of mid-Holocene drying over northern China: Response to a southeastward retreat of the East Asian Monsoon

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[1] To determine the timing and spatial distribution of mid-Holocene drying over northern China, the mineralogical and oxygen isotopic composition of authigenic carbonate from a closed lake at Bayanchagan, southern Inner Mongolia, were measured. Further analysis and synthesis of the spatial geological data were performed. Results from Lake Bayanchagan show a significant drying at 6000 calendar years (cal years) B.P., indicated by dolomite precipitation and a striking rise in $\delta^{18}O$ values. The synthesis of spatial data reveals a zonal distribution for timing of drying over northern China in the mid-Holocene, which began at 9000–7000 cal years B.P. in deserts of north-central China. At 7000–5500 cal years B.P., drying extended into the desert-steppe transitional zone and at ~4500 cal years B.P. into northeastern and south-central China. This pattern indicates that the east Asian summer monsoon has significantly retreated southeastward since the mid-Holocene, which may be related to orbitally induced Northern Hemisphere insolation changes. A retreat of ~400–550 km is inferred for the front of the summer monsoon from 6500 to 4500 cal years B.P.

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## 1. Introduction

[2] The climate of China is mainly controlled by the East Asian Monsoon, which comprises two seasonally alternating atmospheric circulations. In winter, a dry-cold air mass from Siberia leads to a cold and dry climate, while in summer, a southeast monsoon transports heat and moisture inland from the low-latitude oceans, with a gradient of decreasing rainfall going from the southeast to the northwest (Figure 1). The steep inland precipitation gradient is observed in the northern transitional zone between the desert and steppe landscapes, which are particularly sensitive to monsoon precipitation changes.

[3] Previous studies have shown that a dry climate followed early Holocene humid conditions in northern China [e.g., *An et al.*, 2000; *Zhou et al.*, 2001, 2002; *C.-T. A. Chen et al.*, 2003; *Jiang et al.*, 2006]. However, the initiation of mid-Holocene drying varies between sites in northern China, starting between 9000 and 4000 years ago [e.g., *An et al.*, 1993; *Chen et al.*, 1999; *Zhou et al.*, 2001, 2002; *Liu et al.*, 2002; *Shi et al.*, 2002; *C.-T. A. Chen et al.*, 2003; *F.-H. Chen et al.*, 2003; *An et al.*, 2003; *Li et al.*, 2003; *He et al.*, 2004; *Jiang et al.*, 2006]. Clearly, the spatial distribution of the mid-Holocene drying event needs to be identified, not only because this information is crucial to an understanding of climate mechanisms and as a means of defining model boundary conditions, but also because dry events in northern China played a significant role in the collapse and substitution of Chinese Neolithic cultures [*Wu and Liu*, 2004]. Solutions to this problem largely depend on more precisely dated, high-resolution geological records in climatically sensitive regions of northern China, and further analysis and synthesis of the published data.

[4] In this study, we first present mineralogical and $\delta^{18}O$ records of authigenic carbonate from a basin-closed lake at Bayanchagan (BY), southern Inner Mongolia. This site is situated at the current northern edge of the summer monsoon [*Gao*, 1962], in one of the key regions for reconstructing East Asian Monsoon history. Precise age-constrained, high-resolution paleoclimatic records throughout northern China have been collected and compiled to identify the spatial pattern of mid-Holocene drying and its relationship with the East Asian Monsoon.

## 2. Material and Methods

[5] Lake Bayanchagan (115.21°E, 41.65°N; 1355 m asl) was a ~15 $km^2$ closed-basin lake in 1959. Today, it is almost completely dry because of anthropogenic water use, with only small patches of shallow water maintained by summer rain. A 1.8 m trench was cut (core BY, Figure 2a) into the center of Lake Bayanchagan sediment. The chronology of the core is constrained by seven radiocarbon dates of bulk organic carbon (Figure 2b) [*Jiang et al.*, 2006]. Assuming a relatively constant proportion of aquatic to nonaquatic organic carbon throughout the core, the reservoir effect is ~570 years for Lake Bayanchagan [*Jiang et al.*, 2006].

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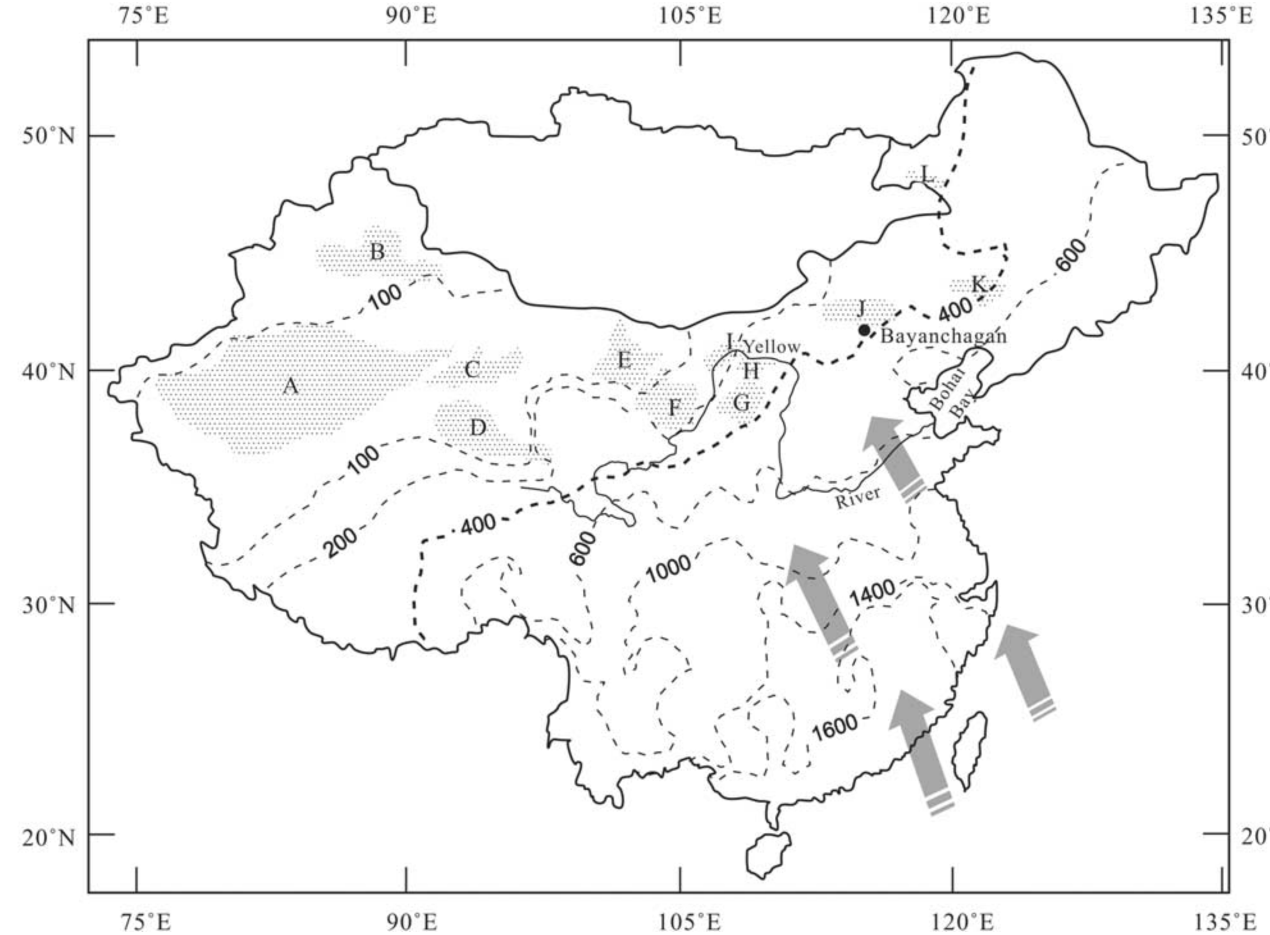


**Figure 1.** Map showing the location of Lake Bayanchagan, deserts and distribution of mean annual precipitation (mm) in China. Precipitation data are from the National Climate Centre of China Meteorological Administration. Deserts are marked as follows: A, Taklimakan; B, Gurbantunggut (Junggar); C, Kumtag; D, Qaidam; E, Badain Jaran; F, Tengger; G, Mu Us; H, Hobq; I, Ulan Buh; J, Otindag; K, Horqin; and L, Hulun Buir. The arrows indicate the advance of the east Asian summer monsoonal rainfall belt.

[6] Carbonate content and $\delta^{18}O$ values of carbonate were determined at 2 cm intervals. Prior to $CO_2$ extraction, all samples were sieved and the <40 $\mu$m fraction collected for further analysis, as the carbonate in the <40 $\mu$m fraction of lake sediment was considered to be authigenic origin [*Fontes et al.*, 1996]. $\delta^{18}O$ values were determined using a Finnigan MAT252 mass spectrometer and are reported in per mil units (‰) relative to the Vienna Peedee belemnite (VPDB) standard. Replicate analyses ($n$ = 5) show that this procedure yields a precision of better than ±0.2‰. X-ray diffraction (XRD, RINT2000 Wide-angle goniometer), and scanning electron microscopy (SEM, LEO 1450VP) were performed to identify the composition and morphology of the authigenic carbonate.

## 3. Results and Discussion

### 3.1. Mineralogical and Oxygen Isotopic Composition of the Authigenic Carbonate From BY Core

[7] Carbonate content is abundant in the BY core (Figure 2c), increasing from 8–40% before 10,800 calendar years (cal years) B.P. to 40–55% between 10,800 and 7200 cal years B.P., and subsequently decreasing gradually from 55% to 34% from 7200 cal years B.P. to the present. XRD results show that calcite predominates in the carbonate fraction between 12,400 and 6000 cal years B.P. (depth 180–48 cm), and dolomite occurs after 6000 cal years B.P. (depth 47 cm) (Figure 3). Using SEM, the calcite fraction consists of small (1–5 $\mu$m long), flaky and lenticular idiomorphic crystals. Dolomite occurs as a knobbly coating (<1 $\mu$m) on the surface of feldspar (Figure 4). The morphological features of calcite and dolomite indicate rapid carbonate precipitation [*Fontes et al.*, 1996].

[8] From 12,400 to 10,800 cal years B.P., all $\delta^{18}O$ values of authigenic carbonate are between −3 and −1‰ VPDB (Figure 2d). From 10,800 to 7800 cal years B.P., the oxygen isotope value decreases with lowest $\delta^{18}O$ values (averaging −7.0‰) observed between 7800 and 7200 cal years B.P. From 7200 to 6000 cal years B.P., the oxygen isotope value gradually increases to between −6.2 and −5.3‰. This trend is followed by a rapid increase of 4.1‰ in $\delta^{18}O$ values from 6000 to 4800 cal years B.P. After 4800 cal years B.P., the $\delta^{18}O$ values show small variations between −1.5 and −0.2‰.

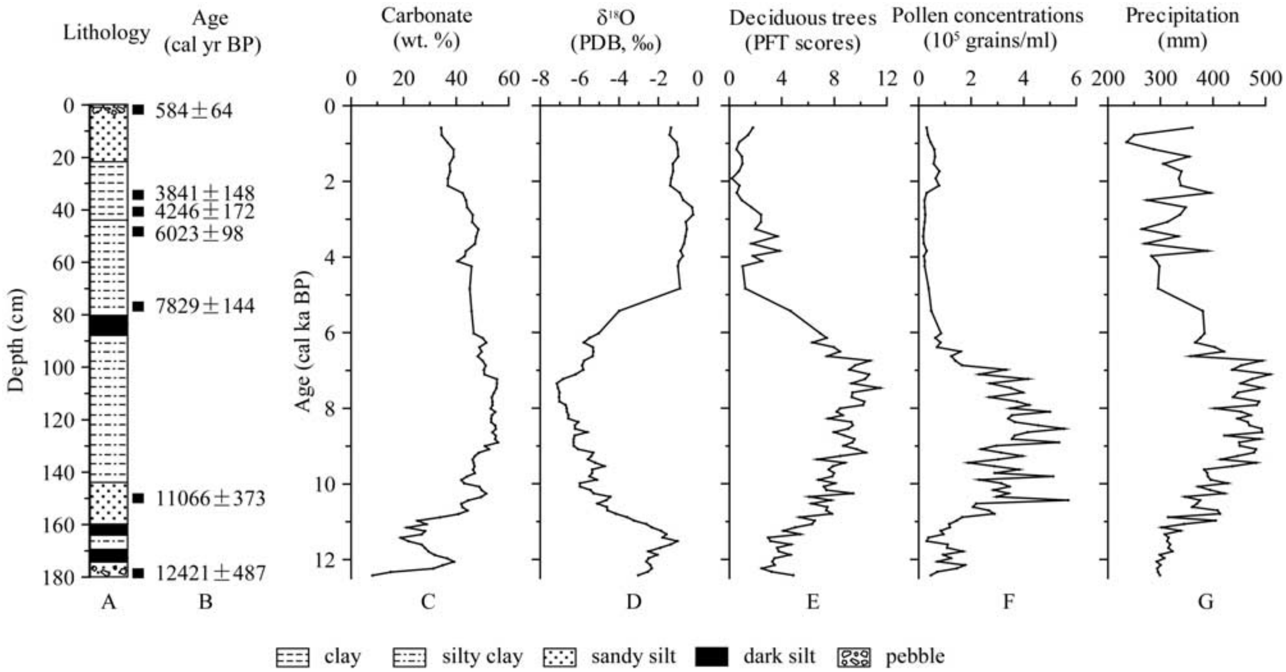


**Figure 2.** (a) Lithology, (b) chronology, (c) carbonate contents, (d) $\delta^{18}$O values, (e) PFT scores of deciduous trees, (f) pollen concentrations, and (g) estimated mean annual precipitation at core BY. Figures 2a, 2b and 2e–2g are from *Jiang et al.* [2006].

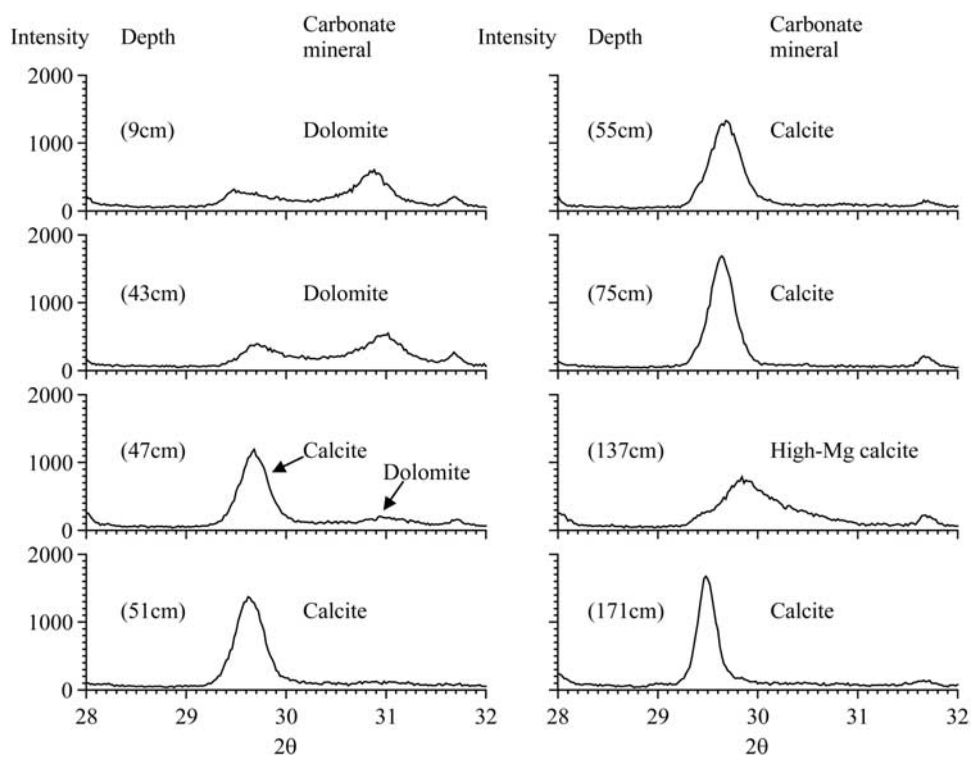


**Figure 3.** X-ray diffractogram (28° to 32° 2$\theta$) for the <40 $\mu$m fraction from the core BY. Below depth of 48 cm, the major peak occurs at 3.01Å (29.62° 2$\theta$), reflecting dominance of calcite. The flat peak at depth of 137 cm represents a relatively high content of magnesium in calcite. At depth of 47 cm, the small increase in the intensity at 2.89Å (30.96° 2$\theta$) indicates the presence of dolomite. Above a depth of 43 cm, the major peak occurs at 2.89Å (30.96° 2$\theta$), indicating an increase of dolomite [*Vasconcelos et al.*, 1995].

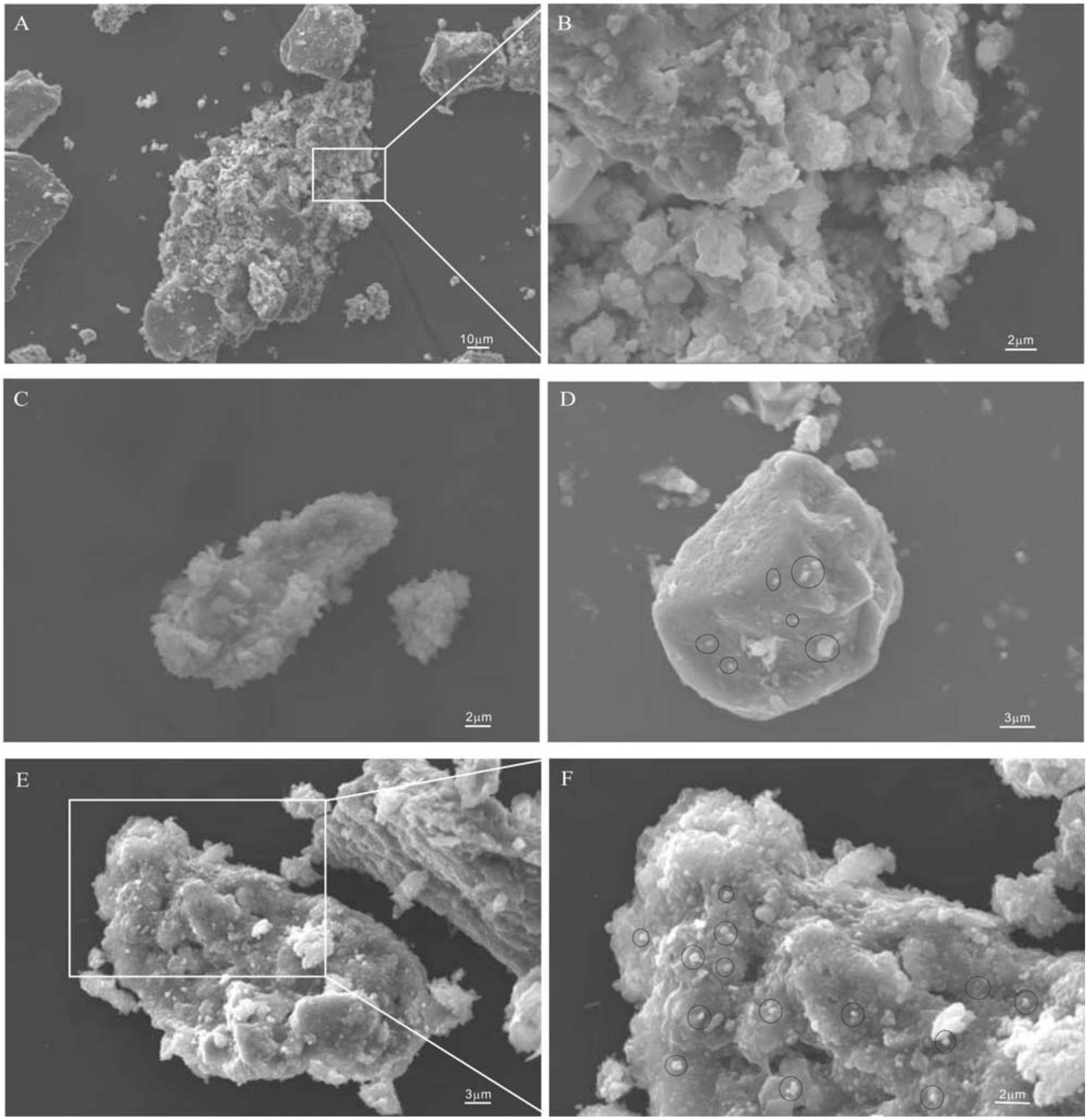


**Figure 4.** SEM photographs of <40 $\mu$m fraction from BY core showing (a) an overview of calcite aggregates on the surface of feldspar, (b) an enlarged picture of the rectangle in Figure 4a, (c) two calcite aggregates, (d) calcite crystals precipitated on the surface of quartz as marked by circles, (e) knobbly dolomite coating on the surface of feldspar, and (f) an enlarged picture of the rectangle in Figure 4e. The dolomite marked by circles has morphological characteristics similar to those of dolomite precipitated in experiments [*Vasconcelos et al.*, 1995].

### 3.2. Mid-Holocene Drying Inferred From Dolomite Precipitation and $\delta^{18}$O Record in BY Core

[9] In lake environments, carbonate phases are controlled by salinities and Mg/Ca ratios of lake water [*Müller et al.*, 1972]. As salinity increases, low-Mg calcite precipitates first, followed by high-Mg calcite, aragonite, and finally dolomite. In core BY, the presence of dolomite around 6000 cal years B.P. provides depositional evidence for salinization, indicative of increasing aridity since the mid-Holocene in southern Inner Mongolia.

[10] Increased aridity beginning in the mid-Holocene is supported by the $\delta^{18}$O record of the authigenic carbonate. $\delta^{18}$O values of authigenic carbonate are controlled by temperature and the oxygen isotope value of lake water [e.g., *Fontes et al.*, 1996], and carbonate phases [*Land*, 1980; *McKenzie*, 1981]. Previous studies have shown that dolomite is enriched in $^{18}$O relative to cogenetic calcite by

**Table 1.** Paleoclimatic Records Collected in This Study[a]

| Site Number | Site | Record | Proxies | Reference |
|---|---|---|---|---|
| 1 | Hongshui River | fluvial-lacustrine | pollen, TOC, $\delta^{18}O$, $\delta^{13}C$, $CaCO_3$, CE | *Zhang et al.* [2000] |
| 2 | Yema Lake | lake | grain size, carbonate | *Shi et al.* [2002] |
| 3 | Alashan Plateau | lakes | pollen | *F.-H. Chen et al.* [2003] |
| 4 | Yanhaizi Lake | lake | grain size, MS, TOC, C/N | *C.-T. A. Chen et al.* [2003] |
| 5 | Midiwan | loess-soil-sand | pollen, $\delta^{13}C$ | *Zhou et al.* [2001], *Li et al.* [2003] |
| 6 | Loess Plateau | loess-soil | pollen, $\delta^{13}C$, TOC | *Zhou et al.* [2001] |
| 7 | Dadiwan | loess-wetland | pollen, mollusk, TOC, grain size | *An et al.* [2003] |
| 8 | Sujiawan | loess-wetland | pollen, mollusk, TOC, grain size | *An et al.* [2003] |
| 9 | Baxie | loess-soil | MS, grain size, TOC, $\delta^{13}C$ | *An et al.* [1993] |
| 10 | Zoige | peat | TOC, gray scale | *Zhou et al.* [2002] |
| 11 | Shanbao Cave | stalagmite | $\delta^{18}O$ | *Shao et al.* [2006] |
| 12 | Mianyang | lake | pollen | *Yang et al.* [1998] |
| 13 | Qidong | Yangtze delta | pollen, TOC, $CaCO_3$ | *Liu et al.* [1992] |
| 14 | Huanghe delta | Huanghe delta | pollen | *Yi et al.* [2003] |
| 15 | Taishizhuang | peat | pollen, $\delta^{18}O$ | *Jin and Liu* [2002], *Tarasov et al.* [2006] |
| 16 | Bayanchagan Lake | lake | pollen, *Pediastrum*, carbonate mineral, $\delta^{18}O$ | *Jiang et al.* [2006], this study |
| 17 | Haoluku | ancient lake | pollen, grain size, LOI, CE | *Liu et al.* [2002] |
| 18 | Xiaoniuchang | ancient lake | pollen, grain size, LOI, CE | *Liu et al.* [2002] |
| 19 | Gushantun | peat | pollen | *Liu* [1989], *Sun et al.* [1991] |
| 20 | Jinchuan | peat | pollen | *Sun et al.* [1991], W. Y. Jiang and T. S. Liu (unpublished data, 2006) |
| 21 | Sandaolaoyefu | peat | pollen | *Yuan and Sun* [1990], *Sun et al.* [1991] |
| 22 | Hulun Lake | lake | pollen, diatom, Ostracode | *Yang et al.* [1995] |

[a]CE, chemical elements; TOC, total organic carbon; LOI, loss on ignition; MS, magnetic susceptibility.

~3.0‰ [*Land*, 1980; *McKenzie*, 1981]. Assuming a maximum percentage of 100% for dolomite after 4800 cal years B.P., then a minimum variation of ~3.0‰ in $\delta^{18}O$ values is expected for cogenetic calcite during the time interval 7200–4800 cal years B.P. (Figure 2d). If temperature were the major factor controlling carbonate $\delta^{18}O$ values, then the minimum variation of ~3.0‰ would have resulted from at least a 14°C temperature change [*Kim and O'Neil*, 1997]. Because such a large change in temperature is unlikely, we suggest that the $\delta^{18}O$ values of authigenic carbonate are mainly controlled by the oxygen isotope value of lake water, i.e., the balance between precipitation and evaporation [*Leng and Marshall*, 2004]. Increased monsoonal precipitation produces lower water $\delta^{18}O$ values while decreased precipitation generates higher $\delta^{18}O$ values, as demonstrated in previous studies [e.g., *Wei and Gasse*, 1999]. Evaporation of lake water results in higher $\delta^{18}O$ values in authigenic carbonate [*Leng and Marshall*, 2004]. Our $\delta^{18}O$ values suggest that precipitation gradually decreased from 7200 to 6000 cal years B.P., and then decreased more rapidly from 6000 to 4800 cal years B.P., indicating progressive drying between 7200 and 4800 cal years B.P. Here we consider the presence of dolomite at 6000 cal years B.P. as marking the beginning of the mid-Holocene drying event at core BY. The mineralogical and oxygen isotopic changes are consistent with pollen data and mean annual precipitation estimated from the same core (Figures 2e–2g) [*Jiang et al.*, 2006].

### 3.3. Timing and Spatial Distribution of Mid-Holocene Drying Over Northern China and Its Implications for the East Asian Monsoon Changes

[11] Paleoclimatic records that range from Tengger Desert to Bohai Bay (Figure 1) over northern China (Table 1, selection criteria are the same as *An et al.* [2006]) have been collected and compiled from lacustrine deposits, peat, loess and stalagmites. Most sites are located in the transitional area between desert and steppe, a climatically sensitive zone, approximately paralleling the 400 mm isohyet of mean annual precipitation.

[12] We identified mid-Holocene dry events through vegetation changes, the termination of soil development and the decline in lake levels [e.g., *An et al.*, 1993; *Zhou et al.*, 2001; *Liu et al.*, 2002; *Shi et al.*, 2002; *F.-H. Chen et al.*, 2003], as these changes are largely related to precipitation [*An et al.*, 2000]. For vegetation records from arid and semiarid regions, the date of increase in desert vegetation (e.g., *Ephedra* and Chenopodiaceae) was taken as the time of drying [e.g., *Zhang et al.*, 2000], while the date of decrease in broadleaved deciduous/evergreen trees and/or increase in steppe vegetation was selected for those from semihumid and humid regions [e.g., *Liu et al.*, 1992; *Yang et al.*, 1998; *Liu et al.*, 2002; *An et al.*, 2003; *Tarasov et al.*, 2006].

[13] Humidity reconstruction in deserts was obtained mainly from pollen and geochemical records. A fluvial-lacustrine record from the terrace of the Hongshui River in the Tengger Desert (site 1, Figure 5) [*Zhang et al.*, 2000] suggests a significant increase in desert vegetation (*Ephedra* and Chenopodiaceae) at 7500 cal years B.P., reflecting enhanced aridity. Lake records indicate a severe drought starting between 7200 and 7000 cal years B.P. in the Tengger Desert (sites 2 and 3, Figure 5) [*Shi et al.*, 2002; *F.-H. Chen et al.*, 2003], and 8800 cal years B.P. in the Hobq Desert (site 4, Figure 5) [*C.-T. A. Chen et al.*, 2003]. A drought that started at 8300 cal years B.P. is also indicated by a pollen record near the margin of Mu Us Desert (site 5, Figure 5) [*Li et al.*, 2003].

[14] In the desert-steppe transitional zone, paleoclimatic records mainly come from loess and lake sediments. Loess deposits have long been regarded as one of the most important archives of climatic evolution in China during

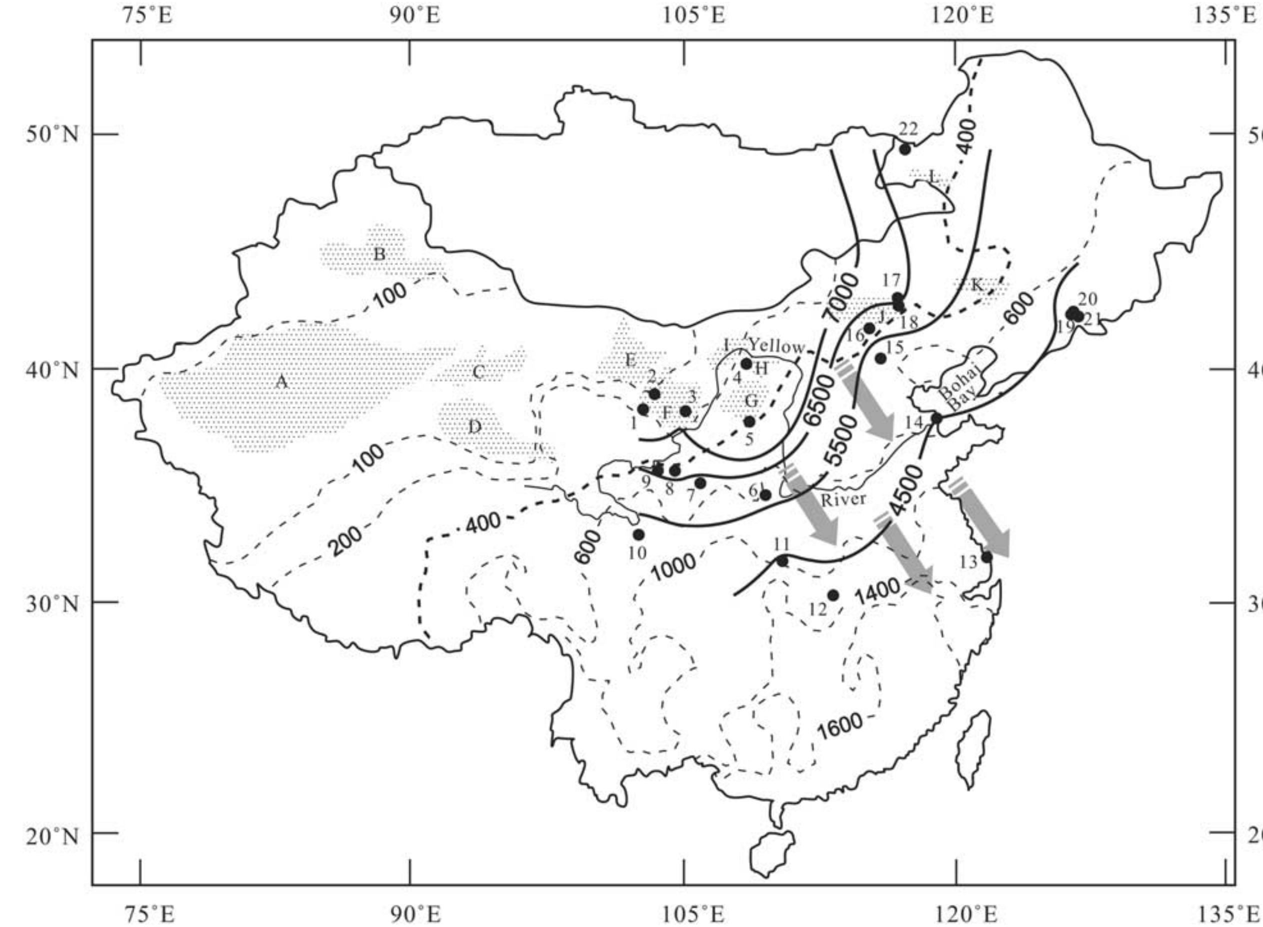


**Figure 5.** Map showing the timing (solid lines, cal years B.P.) and spatial distribution of mid-Holocene drying over northern China. Sites are numbered according to Table 1. Dashed lines are isohyets of mean annual precipitation (mm). Deserts are marked as follows: A, Taklimakan; B, Gurbantunggut (Junggar); C, Kumtag; D, Qaidam; E, Badain Jaran; F, Tengger; G, Mu Us; H, Hobq; I, Ulan Buh; J, Otindag; K, Horqin; and L, Hulun Buir. The arrows indicate the retreat of the summer monsoonal rainfall belt.

the past 2.6 Ma [*Liu and Ding*, 1998]. Loess beds were deposited when climate was arid and cold, whereas soils developed under relatively warm and humid conditions [*Liu*, 1985; *Kukla*, 1987; *Liu and Ding*, 1998]. Previous studies have shown that development of Holocene soil started at 11,500–10,000 cal years B.P., and terminated at 6800–5700 cal years B.P. (sites 6 and 9, Figure 5) [*An et al.*, 1993; *Zhou et al.*, 2001]. Pollen records from lake sediment demonstrate that vegetation type and cover deteriorated at the same time as soil termination, characterized by decreases in pollen concentrations (sites 16 and 22, Figure 5) [*Yang et al.*, 1995; *Jiang et al.*, 2006] and arboreal pollen percentages (sites 16–18, Figure 5) [*Liu et al.*, 2002; *Jiang et al.*, 2006]. The good agreement in timing between soil termination and vegetation changes suggests a mid-Holocene drying at 7000–5500 cal years B.P. in the desert-steppe transitional zone.

[15] There are few paleoclimatic records, with the exception of pollen and stalagmites, available for northeastern and south-central China. Broadleaved deciduous forest was replaced by a mixed forest of deciduous and coniferous trees at 4400–4200 cal years B.P. in northeastern China (sites 19–21, Figure 5) [*Liu*, 1989; *Yuan and Sun*, 1990; *Sun et al.*, 1991; W. Y. Jiang et al., unpublished data, 2006]. An increase in steppe vegetation occurred at 4800 cal years B.P. at the Taishizhuang peat bog (site 15, Figure 5) [*Jin and Liu*, 2002; *Tarasov et al.*, 2006] and 4500 cal years B.P. at the delta of the Yellow River (site 14, Figure 5) [*Yi et al.*, 2003]. Pollen records from south-central China have shown that a decrease in subtropical broadleaved evergreen trees, and an increase in trees with greater moisture tolerances, such as the deciduous genera *Quercus*, *Corylus* and the coniferous genus *Pinus*, occurred at 4300–4200 cal years B.P. (sites 12 and 13, Figure 5) [*Liu et al.*, 1992; *Yang et al.*, 1998]. A high-resolution oxygen-isotope record of a precisely dated stalagmite from Shanbao Cave (site 11, Figure 5) reveals an abrupt decrease in monsoonal precipitation at 4400 cal years B.P. [*Shao et al.*, 2006]. These records suggest a dry, cold climate after 4800–4200 cal years B.P.

[16] On the basis of the above geological records, together with the BY record, we compiled a contour map showing the spatial changes in timing of mid-Holocene drying over northern China using Surfer software. It shows a clear zonal distribution from northwest to southeast over northern China (Figure 5). Drying began at 9000–7000 cal years B.P. in the deserts of north-central China, extending into the

transitional zone between desert and steppe at 7000–5500 cal years B.P. and at ~4500 cal years B.P. into northeastern and south-central China. This zonal distribution of drying time in the mid-Holocene roughly parallels the isohyets of mean annual precipitation in northern China (Figure 5), indicating a close relationship with the east Asian summer monsoon.

[17] The summer monsoonal precipitation is produced by the interaction between warm moist southerly air masses and cold northerly airflows from middle and high latitudes. In general, the more northerly the penetration of the rainfall belt into interior continental northern China, the greater is the intensity of the summer monsoon [*Tao and Chen*, 1987]. The Pleistocene witnessed numerous retreat-advance cycles of the east Asian summer monsoon in response to glacial-interglacial cycles [*Liu and Ding*, 1998; *Ding et al.*, 2001, 2002]. For any specific site in northern China, it would usually become drier when the summer monsoon retreated southward. The progressive drying from 7200 to 4800 cal years B.P. recorded in Lake BY is consistent with the progressive weakening of the east Asian summer monsoon. In this context, the zonal distribution of the desiccation suggests that the East Asian Monsoon has significantly retreated southeastward since the mid-Holocene. Furthermore, a ~400–550 km retreat is inferred for the summer monsoon from 6500 to 4500 cal years B.P.

[18] Changes in the intensity of the east Asian summer monsoon during the Holocene may be related to orbitally induced Northern Hemisphere insolation variability [*An et al.*, 2000; *Kutzbach et al.*, 2001; *Wang et al.*, 2005]. According to *Berger and Loutre* [1991], Northern Hemisphere summer solar insolation has been decreasing since the early Holocene. Numerical modeling experiments have shown that a decrease in insolation would have resulted in a weakened monsoon circulation [*Ganopolski et al.*, 1998; *Kutzbach et al.*, 2001]. The forcing mechanism lies in that the cooling of the continental interior gives rise to a decrease in thermal contrast between the east Asian continent and the adjacent ocean, thus leading to a weakened monsoon circulation, i.e., a southeastward retreat of the east Asian summer monsoon.

## 4. Conclusions

[19] In Lake Bayanchagan, precipitation of dolomite started at around 6000 cal years B.P. $\delta^{18}O$ values of authigenic carbonate increased gradually from 7200 to 6000 cal years B.P., and then rapidly from 6000 to 4800 cal years B.P. All these data indicate the initiation of significant drying at 6000 cal years B.P. in southern Inner Mongolia.

[20] Synthesis of the spatial data suggests a zonal pattern for the timing of mid-Holocene drying over northern China. The drying began at 9000–7000 cal years B.P. in the deserts of north-central China, then extended into the desert-steppe transitional zone at 7000–5500 cal years B.P., and at ~4500 cal years B.P. into northeastern and south-central China. This pattern suggests that the east Asian summer monsoon has retreated significantly southeastward since the mid-Holocene, which may be related to orbitally forced Northern Hemisphere insolation variation. A retreat of ~400–550 km is inferred for the front of the summer monsoon from 6500 to 4500 cal years B.P. This information will provide valuable records for validating GCM models for ancient climate in this region.

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