A penultimate glacial monsoon record from Hulu Cave and two-phase glacial terminations

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ABSTRACT

Oxygen isotope records of three stalagmites from Hulu Cave, China, extend the previous high-resolution absolute-dated Hulu Asian Monsoon record from the last to the penultimate glacial and deglacial periods. The penultimate glacial monsoon broadly follows orbitally induced insolation variations and is punctuated by at least 16 millennial-scale events. We confirm a Weak Monsoon Interval between 135.5 ± 1.0 and 129.0 ± 1.0 ka, prior to the abrupt increase in monsoon intensity at Asian Monsoon Termination II. Based on correlations with both marine ice-rafted debris and atmospheric CH_4 records, we demonstrate that most of marine Termination II, the full rise in Antarctic temperature and atmospheric CO_2, and much of the rise in CH_4 occurred within the Weak Monsoon Interval, when the high northern latitudes were probably cold. From these relationships and similar relationships observed for Termination I, we identify a two-phase glacial termination process that was probably driven by orbital forcing in both hemispheres, affecting the atmospheric hydrological cycle, and combined with ice sheet dynamics.

Keywords: Hulu Cave, penultimate glacial, Termination II, Asian Monsoon, speleothem records, Chinese interstadials.

INTRODUCTION

Climate records of the last glacial period (marine isotopic stages [MIS] 4–2) are characterized by rapid oscillations between cold and warm states (Voelker, 2002). These dramatic events were first clearly seen in Greenland ice cores (e.g., Dansgaard et al., 1993), then in other records, including marine sediments and cave deposits. We previously reported high-resolution absolute-dated Asian Monsoon (AM) records over the last glacial period from Hulu and Dongge caves, China, and demonstrated a strong correlation with Greenland climate between 75 and 11 thousand years before present (k.y. B.P.; present = A.D. 1950) (Wang et al., 2001). However, high-resolution records of climate variability during the penultimate glacial period (MIS 6) and subsequent deglacial sequence are few, and include paleoceanographic studies (e.g., McManus et al., 1999; de Abreu et al., 2003). An understanding of MIS 6 climate may help to unravel the underlying dynamics that control glacial climate variability and the causes of glacial terminations (Broecker and van Donk, 1970). We reported a portion of the MIS 6 AM record from Dongge Cave (25°17′N, 108°5′E; 1200 km WSW of Hulu Cave), revealing a Weak Monsoon Interval (WMI) prior to AM Termination II (TII) (Yuan et al., 2004; Kelly et al., 2005). Here we present a new $\delta^{18}O$ record from Hulu Cave that extends our previous record (Wang et al., 2001) back to the penultimate glacial period. The data are extensive enough so that we can characterize AM stadial-interstadials and correlate variations with both the Vostok ice core record and marine oxygen isotope records during the penultimate glacial period.

SAMPLES AND METHODS

Three stalagmite samples, MSP, MSX, and MSH, were collected from ~31 to 39 m depth in Hulu Cave, eastern China (32°30′N, 119°10′E) (GSA Data Repository Fig. DR1[1]). Stalagmites are 6–35 cm in diameter and 13–40 cm long. Sector inductively coupled plasma–mass spectroscopy ^{230}Th dating procedures were similar to those described in Kelly et al. (2005). Dating results are listed in Table DR1 (18 for MSP, 12 for MSX, and 12 for MSH), covering the majority of MIS 6 (178–162 k.y. B.P. and 155–129 k.y. B.P.). All dates are in stratigraphic order with typical analytical errors of ~1 k.y. We established chronologies using linear interpolation between dates. Subsampling and stable isotope measurements are similar to those described in Yuan et al. (2004). Stable isotope ratios were measured in 1051 samples (307 for MSX, 400 for MSP, and 344 for MSH), giving a mean resolution of ~83, 55, and 48 yr per sample for MSX, MSP, and MSH, respectively (Table DR2). Local meteorological information was presented in Yuan et al. (2004, their supplemental table).

RESULTS AND DISCUSSION

Test for Equilibrium Deposition

The $\delta^{18}O$ records of MSX and MSP are virtually identical over the contemporaneous growth period 150–133 k.y. B.P., providing a replication test for equilibrium deposition (Fig. 1). No correlation is found between $\delta^{18}O$ and $\delta^{13}C$ (average slope = −0.086 and r^2 = 0.017), suggesting equilibrium deposition (Hendy, 1971). A prolonged period of heavy $\delta^{18}O$ values between 135.5 and 129 k.y. B.P. and the timing of AM TII at 129.0 ± 1.7 k.y. B.P. in the Hulu record also replicate events in the Dongge record (Yuan et al., 2004; Kelly et al., 2005) (Fig. 2), consistent with equilibrium deposition and indicating that the events recorded here were at least regional in extent. The slight discrepancy between MSX and MSP may indicate minor kinetic effects with magnitudes that are small compared to the overall climatic signal. Thus, our records largely reflect changes in the $\delta^{18}O$ of meteoric water and cave temperature. Furthermore, as temperature effects are likely small (Wang et al., 2001; Johnson et al., 2006) compared to the amplitude of our record, changes in cave $\delta^{18}O$ mainly reflect changes in the $\delta^{18}O$ of meteoric precipitation.

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[1]GSA Data Repository item 2006040, Figures DR1–DR5 and Tables DR1–DR2, is available online at www.geosociety.org/pubs/ft2006.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

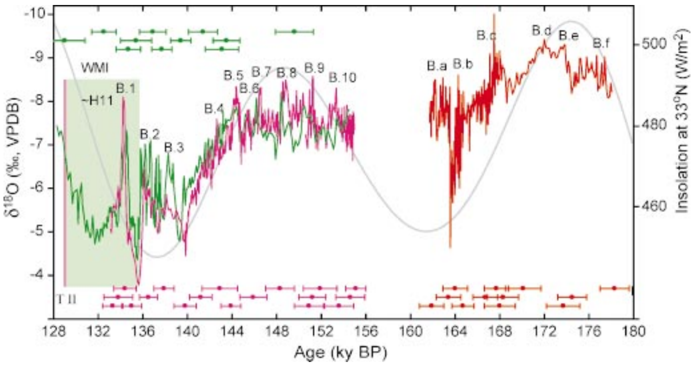

Figure 1. δ^{18}O time series during penultimate glacial period for stalagmites MSX (green), MSP (pink), and MSH (red). Error bars are color coded by sample and indicate ^{230}Th dates and 2σ errors. δ^{18}O scale is reversed. Summer insolation at 33°N (gray) is integrated over June, July, and August (Berger, 1978). Pink and light green vertical bars indicate Asian Monsoon Termination II (TII) ca. 129 k.y. B.P. and Weak Monsoon Interval (WMI) between 129 and 135.5 k.y. B.P. Sixteen millennial-scale Chinese interstadials (CIS) have been identified, including 10 between 134 and 155 k.y. B.P. (numbered B.1–B.10) and 6 between 162 and 178 k.y. B.P. (temporarily lettered B.a–B.f). This notation applies to CIS events during penultimate glacial period, while CIS events during last glacial period are numbered beginning with A.1. H—Heinrich Event; VPDB—Vienna Peedee belemnite.

AM Variation and the Stalagmite δ^{18}O Record

AM history has been described as an alternation between the dry-cold northwesterly winter monsoon and the warm-humid southeasterly summer monsoon systems (An, 2000). The winter monsoon prevails in autumn, winter, and early spring, resulting in limited precipitation largely from proximal moisture sources. Onset of the summer monsoon occurs in late May–early June as the difference between land and ocean temperatures increases, driving strong atmospheric circulation and abundant precipitation from distal sources (Fig. DR2; see footnote 1). The δ^{18}O of AM precipitation is most strongly affected by the fraction of water vapor removed during transfer from the tropical Indo-Pacific to China (Yuan et al., 2004). The larger the fraction removed, the lighter the precipitation. Therefore, δ^{18}O of precipitation in southeastern China changes from heavy to light values at the abrupt onset of the summer monsoon, becoming progressively heavier with its retreat during autumn (Fig. DR3). The seasonal cycle is a plausible analog for the millennial-scale strong AM events (defined here as Chinese interstadials, CIS) and weak AM events. Thus, the δ^{18}O of precipitation may be governed by relative summer to winter monsoon intensities, making the Hulu record a proxy for AM precipitation.

Chinese Interstadials

Similar to the Hulu record of the last glaciation (Wang et al., 2001), the MIS 6 record closely follows Northern Hemisphere insolation (Fig. 1), indicating that orbital forcing plays an important role in driving the AM. At least 16 millennial-scale (strong AM) events punctuate the orbital pattern during MIS 6 (CIS B.1–B.10 and CIS B.a–B.f; lowercase letters in the latter until the record between CIS B.10 and CIS B.a is determined; Fig. 1), similar to the last glacial period AM events that correlate with Dansgaard-Oeschger events (Wang et al., 2001). We introduce a new nomenclature here wherein we denote CIS events of the last glacial period as CIS A.1, etc., numbered from youngest to oldest, and we denote the CIS events for the penultimate glacial period as CIS B.1, etc., also numbered from youngest to oldest (Fig. 1; Figure DR5 [see footnote 1]). The average frequency of penultimate glacial CIS events is ~2 k.y., similar to that of their last glacial counterparts. The most striking feature in our record, similar to the Dongge record (Kelly et al., 2005), is the long WMI from 135.5 ± 1.0 to 129.0 ± 1.0 k.y. B.P., interrupted by one interstadial (CIS B.1) at 134.5 ± 1.0 k.y. B.P. (Fig. 2A). An abrupt δ^{18}O shift ca. 129 k.y. B.P. defines AM TII and indicates the end of the WMI.

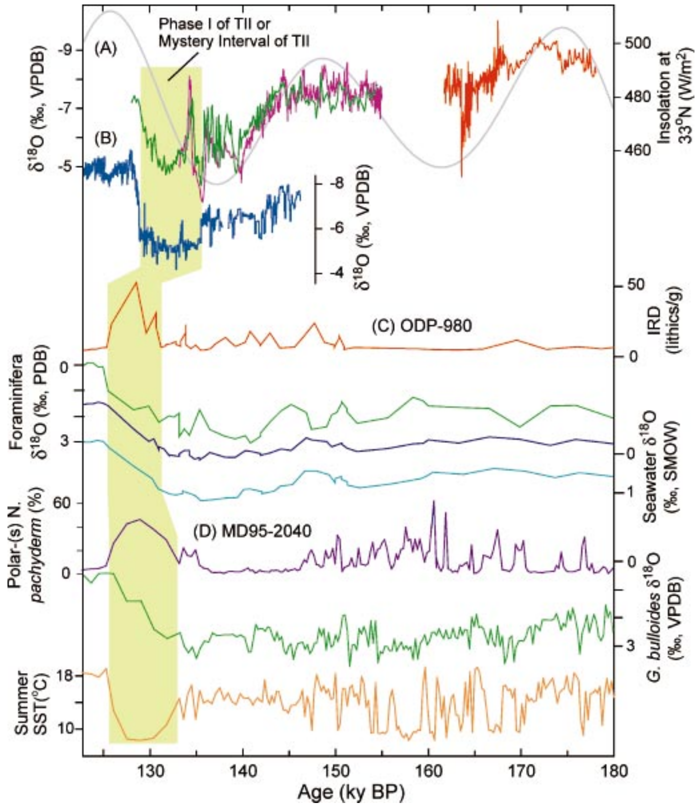

Figure 2. Hulu (A) and Dongge (Kelly et al., 2005) (B) records over penultimate glacial period. Numbers denote Chinese interstadial (CIS) events as in Figure 1. δ^{18}O scale is reversed. C: Ocean Drilling Program (ODP) core 980: ice-rafted debris (IRD) in red (plotted as percent detrital sediment grains >150 mm); foraminifera δ^{18}O of planktic (*Neogloboquadrina pachyderma*, dextral-coiling [*N. pachyderma d.*]) in green; and benthic (*Cibicidoides wuellerstorfi*) in blue; seawater δ^{18}O value in light blue (McManus et al., 1999). D: MD95–2040: δ^{18}O of *Globigerina bulloides* in green; polar-(sinistral) *Neogloboquadrina pachyderma* [(polar-(s) *N. pachyderma*] (%) in purple; and summer sea-surface temperature (SST) in yellow (de Abreu et al., 2003). Vertical green bar denotes phase I or "mystery interval" of Termination II (TII) and shows correlation among Weak Monsoon Interval (WMI), IRD event in core ODP 980, and polar (s) *N. Pachyderma* peak and low SST in MD-95–2040. VPDB—Vienna Peedee belemnite; SMOW—standard mean ocean water.

Two-Phase Terminations

We address the controversial (Winograd et al., 1997) sequence of events around TII with two strategies: correlation of monsoon δ^{18}O to the atmospheric methane record, and the ice-rafted debris (IRD) record. AM δ^{18}O has strong similarities to atmospheric methane around Termination I (TI) (Kelly et al., 2005; Fig. DR4 [see footnote 1]), likely the result of a close link between monsoon precipitation and the extent of low-latitude wetlands (Chappellaz et al., 1993). A similar relationship occurs around TII; the abrupt AM TII δ^{18}O jump has an analog in the methane record ca. 129 k.y. B.P. (Fig. 3). If the abrupt shifts in δ^{18}O and methane correlate, most of the large shifts recorded at Vostok around TII take place during our WMI, when the monsoon was weak and Greenland temperatures were likely cold. Patagonian dust decreases rapidly at the beginning of the WMI. During the WMI, atmospheric CO_2 and Vostok temperature increase through their full glacial-interglacial amplitudes, and atmospheric CH_4 rises through more than half of its glacial-interglacial amplitude (Kelly et al., 2005). AM TII follows these events (Fig. 3).

We further correlate the AM record to the marine record (Fig. 2).

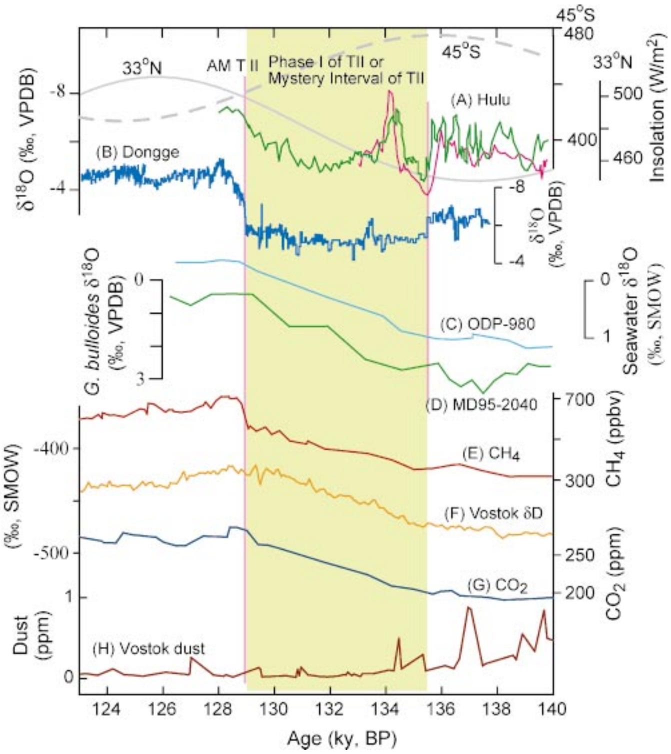

Figure 3. Comparison of (A) Hulu and (B) Dongge records surrounding Termination II (TII). C: Seawater $\delta^{18}O$ value inferred from Ocean Drilling Program (ODP) core 980 (McManus et al., 1999). D: Time series of *Globigerina bulloides* $\delta^{18}O$ in MD95–2040 (de Abreu et al., 2003). Vostok records of (E) CH_4, (F) δD (temperature), (G) CO_2, and (H) dust (Petit et al., 1999). Time scales of ODP core 980 and MD95–2040 have been shifted older by 4 and 3 k.y., respectively, based on correlation from Figure 2. Vostok records are plotted on original GT4 time scale (Petit et al., 1999), which was already nominally consistent with our monsoon CH_4 correlation ca. 129 k.y. B.P. Pink vertical bars denote correlation among Hulu, Vostok, and marine records at onset and end of Weak Monsoon Interval (WMI), shaded green (phase I or "mystery interval" of TII). Insolation curves at 33°N (over June, July, and August; dashed gray line) and at 45°S (December, January, and February; solid gray line) are plotted at top (Berger, 1978). VPDB—Vienna Peedee belemnite; SMOW—standard mean ocean water.

We previously documented a correlation between Weak Monsoon Intervals in the Hulu record and IRD events for the last glacial period (Wang et al., 2001). Correlations among the Hulu record, Greenland ice cores, and deep-sea core MD95–2042 were made for MIS 3 (Shackleton et al., 2004). Here we extend the monsoon-IRD correlation to a key portion of the penultimate glacial period (Fig. DR5; see footnote 1). MIS 6 has only one prominent IRD event at its very end (e.g., Heinrich Event 11 [H11 in Fig. 2] as recorded in Ocean Drilling Program core 980 [McManus et al., 1999]) that correlates with a North Atlantic cold event (Fig. 2; e.g., as recorded in MD95-2042; de Abreu et al., 2003). Our AM record (Figs. 1 and 2) also shows a single prominent weak monsoon event at the end of MIS 6, our WMI between ca. 135.5 ± 1.0 and 129.0 ± 1.0 k.y. B.P. On this basis and the basis of the relationship between IRD events and WMIs for the last glacial period, we correlate the WMI with H11 and the associated cold event. There are no plausible alternate correlations; the nominal duration of the marine cold event is comparable to the duration of the WMI, and all four records in three archives exhibit prolonged extreme values: high IRD concentrations, low summer North Atlantic sea-surface temperatures, high percentage of polar-(sinistral) *Neogloboquadrina pachyderma* [polar-(s) *N. pachyderma*], and heavy cave $\delta^{18}O$ over the interval of the proposed correlation (Fig. 2). Based on this correlation, the $\delta^{18}O$ of seawater ($\delta^{18}O_{sw}$, as inferred from benthic foraminifera) began to shift from glacial values around the beginning of the WMI, and reached values close to interglacial values by the end of the WMI (Fig. 3). The midpoint of the shift in $\delta^{18}O_{sw}$ is ca. 133 k.y. B.P., preceding AM TII by ~4 k.y. This result is consistent within dating error with the ^{230}Th-dated half-height of Bahamas planktonic $\delta^{18}O$ (135 ± 2.5 k.y. B.P.; Henderson and Slowey, 2000). Therefore, we conclude that most of TII sea-level rise occurred during the WMI, while Greenland temperatures were likely cold. However, the last increment of sea-level rise to interglacial values likely occurred at about the time of AM TII (Stein et al., 1993).

The events around TII can be divided into two phases. Phase I covers the several millennia of the WMI (135.5–129 k.y. B.P.), starting with diminishing Patagonian dust and including the majority of sea-level rise, full atmospheric CO_2 and Antarctic temperature rise, and the initial half of the atmospheric CH_4 rise. Phase II (129 k.y. B.P. and immediately thereafter) includes the rapid shift in monsoon $\delta^{18}O$ (AM TII), the final jump in CH_4, inferred abrupt Greenland temperature rise, and the last bit of sea-level rise. In retrospect, we can identify a similar pattern in TI (Fig. DR4; see footnote 1); however, during TI, a smaller proportion of the shift toward interglacial conditions takes place in phase I (ca. 17.5–14.65 k.y. B.P.) compared to phase II (14.65 to ca. 7 k.y. B.P.). Nevertheless, during phase I of TI, sea level rose a full quarter of its glacial-interglacial amplitude over several millennia, there was a Heinrich Event (H1), Greenland temperatures were low, the AM was weak, and both CH_4 and CO_2 rose significantly. Phase I of TI was previously and independently identified by Denton et al. (2005), starting with evidence from European glacial deposits, from which the authors inferred seasonality. Denton et al. referred to this interval as the "mystery interval," a term that we adopt as an alternative to our phase I term for both TI and TII (Figs. 2 and 3; Fig. DR4 [see footnote 1]). Our inferred correlation between Antarctic temperatures and sea level during phase I of both terminations is consistent with observations made by others who studied sea level for the last glacial period (Shackleton et al., 2000; Siddall et al., 2003).

Two-Phase Termination Mechanism

As the phase II portions of both TI and TII coincide with times of high Northern Hemisphere summer insolation, they can broadly be explained by some form of the traditional Milankovitch view of direct boreal insolation forcing. However, phase I of both TI and TII began at times of low boreal summer and high austral summer insolation (Fig. 3). Phase I also occurred at times of low Greenland temperature (observed for TI, inferred for TII), surprising because sea level rose during phase I of both TI and TII, presumably due largely to a decrease in Northern Hemisphere ice sheet volume. An important factor in phase I deglaciation may be the fact that ice volume was near maximum values in the millennia immediately before phase I (Fig. 3). Thus, isostatic depression would have been at a maximum at the beginning of phase I, and ice sheet surface elevation would have been relatively low (all other factors being equal), corresponding to relatively warm surface temperatures (see Peltier and Hyde, 1984; Hyde and Peltier, 1985) and with a relatively high proportion of the ice sheets below sea level. The system may have been poised for deglaciation.

The global pattern of insolation at the beginning of phase I (high austral–low boreal summer insolation) could plausibly have triggered northern ice sheet decay via two processes. Recent speleothem data suggest a Northern Hemisphere–Southern Hemisphere antiphasing of precipitation, and perhaps temperature, on precessional and millennial time scales (Wang et al., 2004a, 2004b). During both phase I's, the Northern Hemisphere data indicate cold-dry conditions (our data; Dansgaard et al., 1993; Wang et al., 2001). During Phase I of TI, the Southern Hemisphere data indicate wet (and possibly warm) conditions (Wang et al., 2004a and 2004b). Thus, high austral summer insolation may have generated warm-wet conditions in Patagonia at the beginning of phase I, causing alpine glacial recession and reducing dust and associated Fe flux to the Southern Ocean (Fig. 3). This may have led to increased CO_2 levels through mechanisms described by Broecker and Henderson (1998, and references therein). If so, average global temperatures would likely have risen, perhaps contributing to phase I decay of the Northern Hemisphere ice sheets.

High austral summer–low boreal summer insolation at the beginning of phase I may have been responsible for the observed dry Northern Hemisphere conditions, perhaps limiting the availability of moisture to the ice sheets, contributing to their decay. Moisture availability has been invoked in the opposite sense to explain the rapid buildup of the ice sheets during the MIS 4 to 5 transition (Ruddiman et al., 1980; Cutler et al., 2003). If dry conditions in the Northern Hemisphere correlate with Southern Hemisphere warmth, the moisture availability idea explains not only our phase I correlations between Antarctic temperature and sea level, but also the sea level–Antarctic temperature correlations for MIS 3 and 4 (Shackleton et al., 2000; Siddall et al., 2003).

If CO_2 rise and reduced moisture transport to the ice sheets caused initial ice sheet decay, continued recession may have been facilitated by the near-maximum depression of the subglacial lithosphere at the beginning of phase I. The length of the WMI is on the order of the time constant for glacio-isostatic adjustments (Cathles, 1975), so lithospheric rebound would have lagged glacial recession throughout phase I, contributing to phase I sea-level rise. In this scenario, a full termination would start with initial high ice volume and high austral summer insolation, followed by high boreal summer insolation. As extremes in insolation occur when eccentricity is high, this mechanism broadly explains why the dominant period in the late Pleistocene glacial-interglacial cycles is that of eccentricity, ~100 k.y.

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