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Variability of the Southwest Monsoon since the Last 25 000 Years and Their Possible Causes: Role of Northern Hemisphere versus Tibetan Plateau

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ABSTRACT: High-precision, clay sediment oxygen and hydrogen isotopes analyses of Pleistocene–Holocene deep-sea sediments from the Bay of Bengal (BOB) are presented for the first time. Our study shows that the major source of sediments in the study area, since the last ~25 000 years, is likely to be the Higher Himalayan crystalline rocks. Further, the study of these stable isotope data displays the variation of southwest monsoon (SWM) in the BOB region since the last ~25 000 years and the cause behind the variation has been interpreted. The $\delta^{18}O$ values of the clay sediments are compared with $\delta^{18}O$ values of the BOB seawater. This comparison shows that the clay sediment $\delta^{18}O$ values of the studied sediment cores temporally vary along with the changes in strength of the SWM. Based on the changes in the clay sediment $\delta^{18}O$ values of the studied sediment, we evaluate the variance in the SWM since the last 25 000 years in the BOB. Our results are consistent with previous work in the region based on other proxies. To evaluate the factors influencing the intensity of the SWM since the last glacial maxima, we conducted comparative analyses of the studied clay sediment $\delta^{18}O$ values with $\delta^{18}O$ values in the Greenland ice cores (GISP2) and Tibetan ice cores (Guliya). The results from this comparative study show that large-scale changes in the intensity of the SWM since 25 000 years are affected by the climate oscillations of the Northern Hemisphere, but rapid and abrupt fluctuations in the SWM seem to be controlled by the amount of snow cover in the Tibetan Plateau.

KEY WORDS: southwest monsoon, Bay of Bengal, Tibetan Plateau, clay, stable isotope.

INTRODUCTION

Variability of the southwest monsoon (SWM) or the Asian summer monsoon affects many facets of the Earth's system. Extreme variations in the SWM can cause drought, flooding, and crop failures and thereby affect lives of almost 30% of the world's population (Webster et al., 1998). The impact of fluctuations in the SWM can be judged by the fact that a repressed SWM is considered to have destroyed the entire Indus

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Valley civilization in the Late Holocene (Staubwasser et al., 2003). Recently, in the mid-1960s, more than 1 million people in India died due to suppression in the SWM for 3 consecutive years (Emergency Disasters Data Base or EM-DAT, 2005). On a global scale, the SWM is an important element of the climatic system and interacts with other components such as North Atlantic thermohaline circulation and amount of snow cover on the Tibetan Plateau (Goswami et al., 2006; Thompson et al., 1997; Overpeck et al., 1996). Therefore, for better paleoclimatic modeling and predicting future patterns of the SWM, it is essential to chart past variations of the SWM and to evaluate the causes behind variations.

Several factors causing changes in the strength of the SWM have been suggested by previous studies. On multimillennial scale (last 20 Ma), SWM intensity variation has been primarily attributed to the rate of uplift and exhumation activities in the Himalayas (Galy et al., 1996; Bouquillon et al., 1990) based on sediment cores retrieved from ODP Leg 116 in the Bay of Bengal (BOB) (Fig. 1). On a comparable timescale, studies from the Arabian Sea suggested that the rates of exhumation and erosion in the Himalayas are coupled with the intensity of the SWM in India (Clift et al., 2008; Theide et al., 2004). On interannual and decadal timescale, Goswami et al. (2006) suggested a coupled ocean-atmosphere phenomenon arguing the existence of a teleconnection between the North Atlantic climate and the SWM. Sirocko et al. (1996), Sinha et al. (2005), and Gupta et al. (2006) have also suggested that the coupling of the Northern Hemisphere climate with SWM intensity existed during the glacial-interglacial period. On the other hand, some studies have argued that the SWM intensity variation during the glacial-interglacial period was controlled by the amount of snow cover over the Tibetan Plateau (Zhao and Moore, 2004). Recently, Rohling et al. (2009) claimed that the correlation between the SWM and the Northern Hemisphere climate is not possible during the glacial-interglacial period. They proposed that the Southern Hemisphere climate changes provide a dominant control on the SWM variability during these times. Clearly, there is a long-standing debate on the factors controlling the SWM intensity during glacial-interglacial periods and

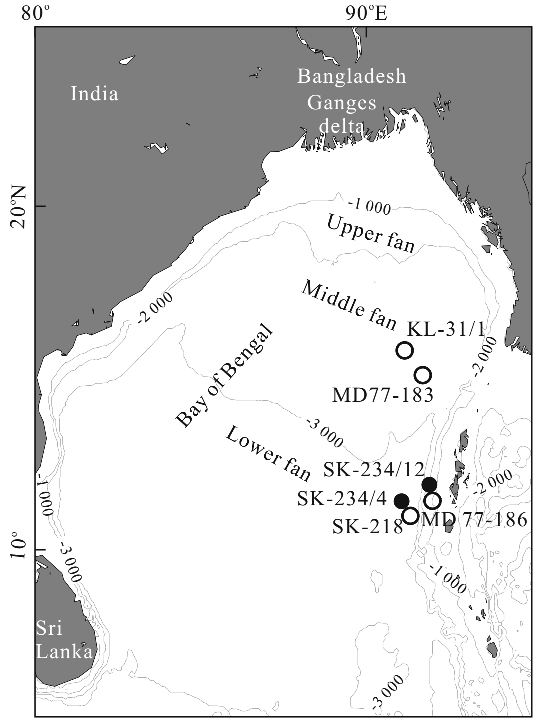

Figure 1. Map showing location of the studied sediment cores (SK-234/4 and SK-234/12). Bathymetric depth contours are in meters. Solid lines demarcate the boundaries between upper, middle, and lower fans. Locations of those cores that are used for age determination are also shown.

both Northern Hemisphere and Southern Hemisphere factors are believed to be playing a major role. The BOB has been the main catchment area for the SWM precipitation as well as for its sedimentation. Therefore, direct evidences from the BOB would facilitate in narrowing down the factors considered to be affecting the SWM intensity. Surprisingly, there are limited observational data available from the BOB (Chauhan et al., 2004; Chauhan, 2003); so far, there is no literature on clay sediment $\delta^{18}O$ and $\delta^{2}H$ composition of Pleistocene–Holocene deep-sea sediments from the BOB.

This article aims to bridge this gap by presenting for the first time $\delta^{18}O$ and $\delta^{2}H$ variations in clay sediments of two Late Quaternary sediment cores of the BOB (Fig. 1). Our objectives are the following: (a) to identify the source(s) of sediments in the BOB since the last glacial maxima (LGM), (b) to study the SWM intensity pattern since the last ~25 000 years, and (c) to evaluate the presence (or absence) of coupling in the intensity of the SWM with the last glacial cycle of the Northern Hemisphere and the Tibetan Plateau.

METHODS AND DATA

Two sediment gravity cores, obtained from the marine expedition SK-234 using ORV/Sagar Kanya, are studied in this article. The two cores, namely SK-234/4 (11º26.51'N and 91º04.37'E) and SK-234/12 (11º55.99'N and 90º55.05'E), were collected at water depths of 3 600 and 4 000 m, respectively, from the Bengal fan (Fig. 1). The sediment cores are of 255 cm (SK-234/4) and 250 cm (SK-234/12) in length. After subsampling at 1 cm interval, the samples were oven dried at 40 ℃, powdered, and homogenized by using agate ball mill. The <2 μm fraction of the sediment was separated by settling velocity principle and clay samples (<2 μm fraction) were dried, powdered, and used for stable isotope measurements.

Oxygen and hydrogen stable isotope analyses were done on 22 such samples at the Cornell Isotope Laboratory. Thermo Delta V isotope ratio mass spectrometer interfaced to a TC/EA was used for the analysis. The standard deviation for the internal standard benzoic acid for $\delta^{18}O$ was 0.18‰ and for $\delta^{2}H$ was 0.88‰. Isotope corrections were performed using two-point normalization (regression) based on their respective international standards (IAEA CO-1 and IAEA CO-8 for $\delta^{18}O$) and a single-point correction for $\delta^{2}H$ using CH-7. The primary reference scale for these isotopes is Vienna Standard Mean Ocean Water (VSMOW). Below 45 cm depth, the general grain size in both the studied sediment cores is <2 μm. This precluded us from absolute dating of the constituent minerals. Therefore, we compiled sedimentation rates at various depths from existing dated sediment cores of the BOB within close proximity of the studied core locations (Fig. 1). For defining the sedimentation rate at different depths of the studied sediment cores, reference core method is applied (Colin et al., 1999). SK-218, SK-31/1, and MD77-186 are used to define sedimentation rate from 3–60, 61–100, and 101–260 cm of depth, respectively. The locations of these reference cores are as follows: SK-218 is at 11º00.44'N, 91º20.10'E; SK-31/1 is at 15º52'N, 91.10ºE; and MD 77-186 is at 11º27.5'N, 92ºE (Fig. 1). SK-218 is nearest to the studied cores and its length is 70 cm (Babu et al., 2010), whereas SK-31/1 core is next proximal dated sediment core with a length of 100 cm (Chauhan, 2003). Subsequent nearest dated core is MD 77-186, which is 310 cm long (Colin et al., 1999). Apparent correlation among the calculated age of the sediment cores and variations in the $\delta^{18}O$ values of the studied sediments as well as of the BOB seawater (Fig. 2) corroborate the time scale used for the cores.

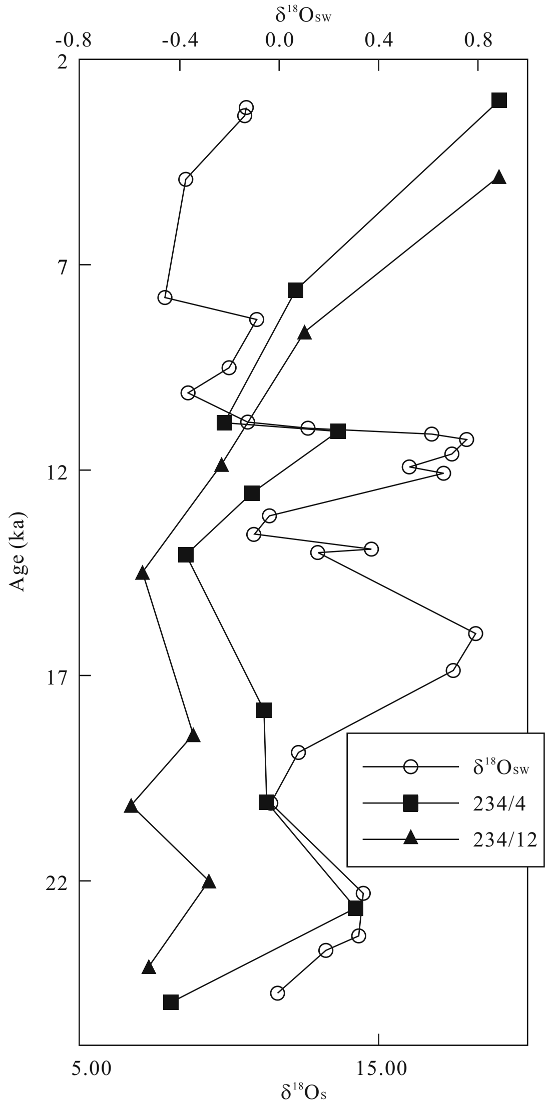

Figure 2. Temporal evolution of the $\delta^{18}O$ isotopic composition of the clay sediments ($\delta^{18}O_s$) cored from the BOB (SK-234/4 and SK-234/12). The intensity of the SWM is traced by $\delta^{18}O$ records of the BOB seawater shown as $\delta^{18}O_{sw}$ (Rashid et al., 2011); $\delta^{18}O$ values are in per mil.

RESULTS

The $\delta^{18}O$ and $\delta^{2}H$ values of the studied sediment cores are presented in Table 1. At ~26 ka B.P., the $\delta^{2}H$

value is -62.24‰ and -59.06‰ in the cores SK-234/4 and SK-234/12, respectively. At ~23 ka B.P., this value decreased to -68.43‰ in the core SK-234/4 and to -63.06‰ in the core SK-234/12. Further, the δ^2H remained similar at ~20 and ~18 ka B.P. in the core SK-234/4. A remarkable low in the values of δ^2H is noticed at ~18 ka B.P., with values of -68.97‰ and -69.34‰ in the cores SK-234/4 and SK-234/12, respectively. On the contrary, both cores display a significant increase in their δ^2H values at ~14 ka B.P., which is -63.83‰ and -62.08‰ in the cores SK-234/4 and SK-234/12, respectively (Table 1). Further, δ^2H value displays a decline at ~12 ka B.P. when it goes down to -68.61‰ in the core SK-234/4 and to -67.17‰ in the core SK-234/12. From ~8 ka B.P. onward, there is a progressive decrease in the δ^2H value in both cores (Table 1), reaching as low as -91.77‰ at ~3ka B.P. in the core SK-234/4 and 79.37‰ at ~5 ka B.P. in the core SK-234/12.

The δ^{18}O in the studied cores at ~26 ka B.P. is 8.03‰ and 9.65‰ in the SK-234/4 and SK-234/12, respectively. At ~23 ka B.P., this value reduces to 7.29‰ in the core SK-234/12, whereas, at ~22 ka B.P., δ^{18}O value raises to 14.23‰ in the core SK-234/4. In the core SK-234/12, at ~21 ka B.P., δ^{18}O value is 9.31‰, which goes down to 6.70‰ at ~20 ka B.P.. Similarly, in the core SK-234/4 also, δ^{18}O value drops down at ~20 ka B.P. when it is 11.25‰. The fluctuations in the δ^{18}O value continue and reach higher value at ~18 ka B.P. and again go low at ~14 ka. During transition from Pleistocene to Holocene, the δ^{18}O values in the clay sediments show a progressive enrichment that continues until ~11 ka B.P.. The enrichment trend is truncated for a short period at ~10 ka B.P., and after that, there is continuous trend of increase in the δ^{18}O values of the sediments until 3 ka B.P..

SEDIMENT SOURCE

We use δ^{18}O and δ^2H isotopic composition of the sediments to identify their source. The δ^{18}O variations of the sediments are in the range of 7‰–12‰ (Table 1). This is slightly away from the 8‰ to 10‰ range of the Himalayan unaltered micas derived from igneous and metamorphic rocks (Derry and France-Lanord, 1996) (Fig. 3). These micas of BOB are reported to be derived from the higher Himalayan crystalline (HHC) rocks, which represent unaltered high-temperature metamorphic sources (Colin et al., 1999; Derry and France-Lanord, 1996). Minor variations in the δ^{18}O values may have been caused due to the presence of weathered components of minerals. Based on isotopic composition of the sediments, we infer that the studied cores are most likely derived from unaltered high-temperature metamorphic sources of the Himalayas and appeared to have undergone minor low-temperature alteration during transport. However, this interpretation has to be taken in the light of the knowledge that there are many other factors, apart from the sediment source, which can affect the δ^{18}O and δ^2H isotopic composition of the sediments.

In the shallower part of the cores, i.e., 15 cm upward, the δ^{18}O values in both cores are ~19‰. These values are similar to the δ^{18}O values of smectite and kaolinite in pedogenic clay from the higher Himalayas (Stern et al., 1997). This may suggest that the shallow-level sediments in both cores have undergone extensive floodplain weathering after their derivation from the source in the higher Himalayas.

RECONSTRUCTION OF THE SWM INTENSITY SINCE LGM

The δ^{18}O value in foraminifera or other biological proxies is primarily controlled by the variations in δ^{18}O composition of the host seawater because their growth is dependent on seawater. Therefore, to comprehend the paleoclimatic information, we compare our clay sediment δ^{18}O values ($\delta^{18}O_s$) with δ^{18}O values of the BOB seawater (referred to as $\delta^{18}O_{sw}$), which has been recently published by Rashid et al. (2011).

The profiles of $\delta^{18}O_s$ and published $\delta^{18}O_{sw}$ show remarkable similarities (Fig. 2). This pronounced similarity and synchronicity in the variations of the $\delta^{18}O_s$ and $\delta^{18}O_{sw}$ are reported for the first time from the BOB and provide an extremely reliable way to estimate paleo-SWM (Fig. 2). These profiles present two important information: (i) the clay sediment $\delta^{18}O_s$ variation is synchronous with δ^{18}O values in the BOB seawater and (ii) the general replication of the profiles of $\delta^{18}O_s$ and $\delta^{18}O_{sw}$ during the entire studied period, that is, 25–3 ka, favors their contemporaneous growth and indicates that the $\delta^{18}O_s$ are affected by neither the

Table 1 Sampling depth, calculated age, $\delta^{18}O$ (normalized $\delta^{18}O$ vs. NSMOW), and $\delta^{2}H$ (normalized $\delta^{2}H$ vs. VSMOW) analyses of the studied sediment cores SK-234/4 and SK-234/12

Sediment core 234/4				Sediment core 234/12			
Depth (cm)	Age (ka)	^{18}O (‰)	^{2}H (‰)	Depth (cm)	Age (ka)	^{18}O (‰)	^{2}H (‰)
5–6	3.00	19.05	-91.77	15–16	4.84	19.05	-79.37
30–31	7.62	12.22	-78.14	25–26	-	No signal	No signal
55–56	10.85	9.82	-71.37	40–41	8.63		
60–61	11.06	13.64	-63.69	65–66	11.85	12.52	-71.34
80–81	12.56	10.75	-68.61	100–102	14.48	9.73	-67.19
100–102	14.06	8.54	-63.83	146–148	18.43	7.08	-62.08
144–146	17.84	11.15	-68.97	166–168	20.15	8.78	-69.34
170–172	20.08	11.25	-68.08	180–182	21.36	6.70	-62.56
200–205	22.66	14.23	-68.43	200–205	23.08	9.31	-64.49
250–255	26.96	8.03	-62.24	245–250	26.95	7.29	-63.06
						9.65	-59.06

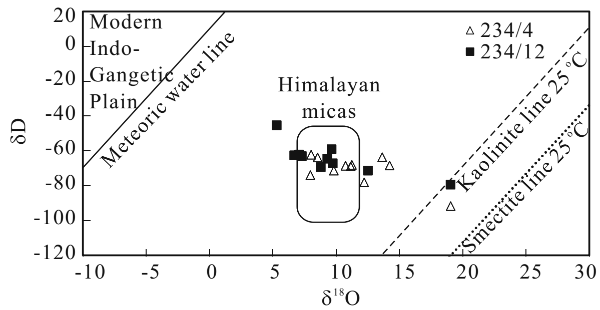

Figure 3. $\delta^{18}O$ (‰) vs. δD ($\delta^{2}H$ in ‰) of clay sediments of the studied cores SK-234/4 and SK-234/12 from the Bengal fan showing that most of the samples are sourced from the HHC rocks. The central box is defined by measurements of HHC outcrop samples (France-Lanord et al., 1988). The fields for kaolinite and smectite in equilibrium and of modern meteoric water of the Indo-Gangetic Plain are from Bouquillon et al., 1990.

water/rock interactions nor other factors. Consequently, $\delta^{18}O_s$ values can be reliably used as paleomonsoon proxy much like the $\delta^{18}O_{sw}$. Because depleted $\delta^{18}O_{sw}$ values suggest a stronger SWM and more sediment discharge from the Himalayas in the BOB (Rashid et al., 2011), it is pertinent from Fig. 2 that the increase in rainfall and sediment discharge resulted in depletion of $\delta^{18}O_s$ values. On the other hand, reduction in rainfall and sediment discharge reflects in enrichment of $\delta^{18}O_s$ values. With this background, variations in the $\delta^{18}O_s$ values of the studied sediments are recognized and interpreted in terms of intensity of the paleo-SWM. The $\delta^{18}O_s$ value shows a progressive enrichment between 25 and 17 ka B.P., with its depletion truncating the overall positive trend at 23 and 20 ka B.P. (Fig. 2). The progressive enrichment of $\delta^{18}O_s$ value during LGM indicates the weakening of SWM intensity, whereas the depleted crests in $\delta^{18}O_s$ values at 23 and 20 ka B.P. (Fig. 2) imply a short and abrupt phase of stronger SWM during an overall dry period. The transition from the last glacial and Holocene at ~14 ka B.P., related to the Bølling-Allerød (B/O) phase of the Northern Hemisphere, is recorded by marked depletion in $\delta^{18}O_s$ values. The depletion at ~14 ka B.P. suggests a strong monsoon in the BOB during the B/O period (Fig. 2). Subsequently, enrichment in $\delta^{18}O_s$ value occurs during 12 and 11 ka B.P., duration considered as the Younger Dryas (YD) period in the Northern Hemisphere paleoclimate. The enriched $\delta^{18}O_s$ values of the BOB sediments during the YD suggests that the SWM was weakest during this period; consequently, lower sediment influx from the GB river system was received in the BOB. A sharp depletion in $\delta^{18}O_s$ values at ~10.8 ka B.P. implies an abrupt, rapid spell of strong SWM in the BOB. From ~8 ka B.P. onward, progressive enrichment of $\delta^{18}O_s$ values indicate progressive weakening of the SWM and commencement of drier climate over the BOB.

Our paleo-SWM reconstruction is highly consistent with other studies in the BOB, Arabian Sea, Andaman Sea, and other continental studies of the Asian summer monsoon (Rashid et al., 2011, 2007; Schmiedl and Mackensen, 2006; Gupta et al., 2005; Fleitmann et al., 2003; Sirocko et al., 1993).

SOURCE OF FLUCTUATION IN THE SWM INTENSITY SINCE LAST 25 KA: NORTHERN HEMISPHERE OR TIBETAN SNOWFIELDS?

The clay sediment $\delta^{18}O$ records in sediment cores of the BOB show near-synchronous fluctuations in SWM intensity as reported by previous studies. Such changes have been attributed to both global and regional processes. Coupling between temperature changes in the Northern Hemisphere, as observed in the Greenland ice core records during the glacial-interglacial stage, and low-latitude monsoonal climate variability implies a link between the two (Schulz et al., 1998). On the other hand, numerous other studies have shown that more snow cover over the Tibetan Plateau can weaken the SWM strength (Zhao and Moore, 2004; Barnett et al., 1988 and references therein). In view of this prevailing ambiguity about the factors affecting the intensity of the SWM, we compare our data with the $\delta^{18}O$ values of the Greenland ice core (GISP2) (representing the Northern Hemisphere paleoclimate changes) and the Guliya ice core (representing changes in paleoclimate of the Tibetan Plateau). The Guliya ice cores are retrieved from Qinghai-Tibetan Plateau (Q-T Plateau), which is also referred as the roof of the Earth (being at the height of 4.5 km) and strongly influences the atmospheric boundary conditions of the tropics (Thompson et al., 1997).

The $\delta^{18}O$ profiles of the ice cores of GISP2 and Guliya, along with the $\delta^{18}O$ profiles of the BOB seawater and the studied sediments, are presented in Fig. 4. The characteristic magnitudes and shapes of most of the peaks and troughs of the GISP2 record are reproduced in the Guliya ice cores and the BOB $\delta^{18}O$ records, both the clay sediments and the seawater. However, comparison of abrupt, small extent, changes in the BOB records with GISP2 and Guliya ice cores shows a different picture.

During the LGM, the $\delta^{18}O$ values in the GISP2 reveal abrupt warm events, called Dansgaard-Oeschger or D-O events (Fig. 4). The D-O events in GISP2 are quite different from the $\delta^{18}O$ value oscillations in the Guliya ice cores (Fig. 4). The $\delta^{18}O$ value oscillation in the BOB records is similar to fluctuations shown by $\delta^{18}O$ values in Guilya ice cores rather than GISP2. One such brief punctuation at 20 ka B.P. is recorded in the $^{18}O_s$ values as well as in $\delta^{18}O_{sw}$ value of the BOB. This low ~20 ka B.P. in $\delta^{18}O$ values of the BOB seawater as well as in the clay sediments indicate an abrupt, short period of strengthened SWM, which should correlate with warming event in the GISP2. However, at this time, the $\delta^{18}O$ values of GISP2 show a cooling low or stadial period (Fig. 4). On the other hand, Guliya ice cores record a warming peak that is consistent with the BOB data (Fig. 4). Therefore, we infer that the abrupt, stronger monsoonal condition at 20 ka B.P. was controlled by regional climatic conditions (warming in Tibet) rather than the climatic variance in the Northern Hemisphere.

Similarly, at ~14 ka B.P., depletion in $\delta^{18}O$ values of the studied BOB sediments and seawater suggests an intense SWM at 14 ka B.P. (Fig. 2), which is in agreement with the results of the other studies (Sinha et al., 2005). Stronger SWM at ~14 ka B.P. represents an abrupt change in a largely weakened SWM during the transition period between LGM and Holocene (Figs. 2 and 4). As shown in Fig. 4, after 14 ka B.P., the $\delta^{18}O_s$ values show a generally rising trend, indicating weakened SWM after 14 ka B.P. until YD. Comparatively, the $\delta^{18}O$ values of the GISP2 show a decrease (cooling) at ~14 ka B.P., which is incoherent with stronger SWM condition in the BOB (Fig. 4). The Guliya ice cores present a progressive rise from 15 to 13.5 ka B.P. followed by a gradual decrease until YD (Fig. 4). This trend is consistent with the SWM reconstruction based on our data, which interpret a strong SWM at 14 ka B.P. followed by a decline until YD.

Based on these observations, we infer that, on larger magnitude, the SWM intensity in the last 25 000 years appears to be coupled with the Northern Hemisphere glacial-interglacial climate. However, this linkage may have been punctuated by brief, decade-to-century long periods of sudden changes in SWM intensity. Such short timescale changes in the SWM are apparently coupled with variation in

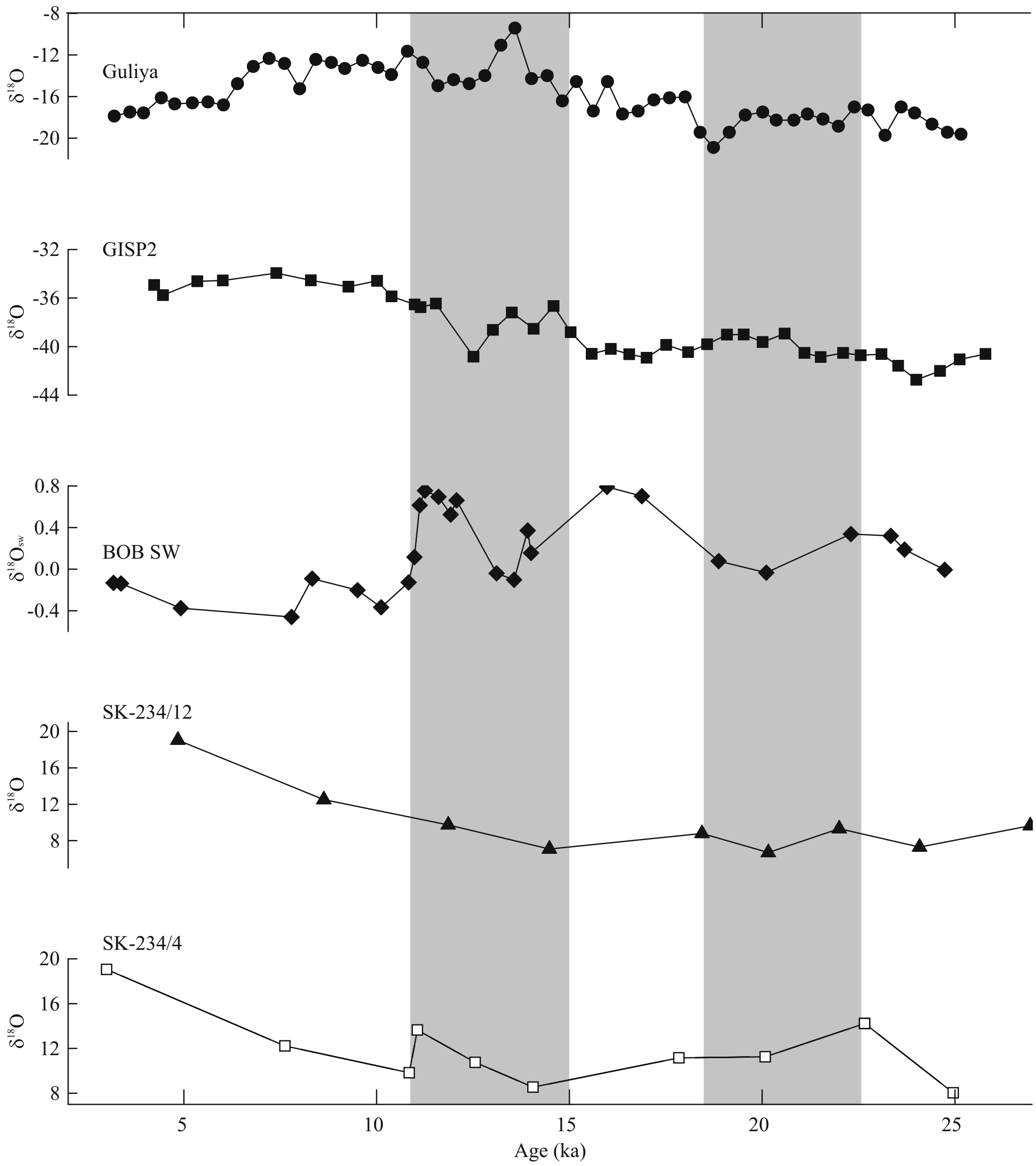

Figure 4. δ18O isotopic evolution of the studied sediments cored from the BOB compared with the Greenland Ice Sheet Project 2 (GISP2) ice core climate record (Stuiver and Grootes, 2000), Q-T Plateau ice core climate record (Thompson et al., 1997), and the intensity of the SWM traced by by δ18O records of the BOB seawater (Rashid et al., 2011).

snowfields over the Tibetan Plateau. Reduced snow cover in Tibetan Plateau would lower the albedo, whereas the exposed soils would increase the radiative heating of the surface. Consequently, these factors would affect the atmospheric boundary conditions in the tropic and change the SWM intensity.

CONCLUSIONS

Our work reports the values of $\delta^{18}O$ and $\delta^{2}H$ from the clay sediments of the deep-sea sediment cores retrieved from the BOB. We demonstrate that the source of glacial-interglacial sediments in the studied part of the BOB is most likely to be the HHC rock. It is shown that $\delta^{18}O$ values of the clay sediment preserve the changes in the intensity of the SWM through glacial-interglacial periods. Evaluation of the factors controlling the SWM intensity through the last 25 000 years shows that, in larger magnitude, SWM is coupled with the GISP2 or temperature changes in the Northern Hemisphere. However, abrupt, shorter time-scale changes of the SWM strength appear to be caused by the changes in the snow cover of the Tibetan Plateau.

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