Modulation and amplification of climatic changes in the Northern Hemisphere by the Indian summer monsoon during the past 80 k.y.

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ABSTRACT

High-frequency suborbital variations (Dansgaard-Oeschger cycles) characterize the climatic history of the Northern Hemisphere as observed in Greenland ice cores, deep-sea sediments of the North Atlantic, the Californian borderland, the Arabian Sea, the South China Sea, and the Chinese loess area. Paleoceanographic data from core KL126 from the Bay of Bengal in combination with data from the other Asian monsoonal areas indicate that the feedback processes involving snow and dust of the Tibetan Plateau vary the summer monsoon capacity to transport moisture into central South Asia and into the atmosphere. We postulate that the summer monsoon initiates, amplifies, and terminates these cycles in the Northern Hemisphere.

Keywords: monsoon, Southeast Asia, Pleistocene, paleoclimate.

INTRODUCTION

The monsoonal system of the Indian Ocean is one of the major atmospheric components of the tropical climate in which there are large seasonal variations with intensive rainfall during summer. During the northern winter, dry cold winds from the Asian continent flow offshore, whereas in central Asia westerly winds flow around the Tibetan Plateau (McGregor and Nieuwolt, 1998) (Fig. 1). With the onset of summer in the Northern Hemisphere, the intense heating of northern India creates a steep pressure gradient between the low onshore and the relatively cool equatorial oceanic high. As a consequence, huge amounts of warm, humid air are transported across the equator and the Arabian Sea into southern China and northern Southeast Asia.

The monsoonal cycle dominates the fluvial runoff of one of the world's largest rivers, the Ganges-Brahmaputra system, which drains most of the Himalayas and the northern Indian subcontinent. During the August peak of the summer monsoon, as much as 50 000 m^3/s of water (Milliman and Meade, 1983) with an average $\delta^{18}O$ content of $-8‰$ (Feng et al., 1999) flows into the Bay of Bengal, lowering the salinity in the bay (Duplessy, 1982).

Here we present paleoceanographic data from a core 8.8 m long taken in the northern Bay of Bengal (SO93–126KL, 19°58.40′N, 90°02.03′E, water depth 1253 m) located 200 km south of the mouth of the Ganges-Brahmaputra (Fig. 1). The core consists of bioturbated silty clay with a carbonate concentration (2%–12%). The prominent volcanic ash layer of the Toba eruption ca. 70 ka (Westgate et al., 1998; Zielinski et al., 1996) is found at 664–666 cm core depth.

METHODS

A record of the isotopic composition of the surface water was established by measuring 10–15 tests of *Globigerinoides ruber* (white) (>315 μm diameter) representing a time resolution of 100–200 yr (Fig. 2). Starting from the isochronous time mark of the Toba ash (70 ka) (Westgate et al., 1998), the $\delta^{18}O$ values of the planktonic foraminifera *Globigerina ruber* (white) were correlated with the Dansgaard-Oeschger (D-O) cycles in the $\delta^{18}O$ of the GISP2 ice core (Fig. 3) (Grootes et al., 1993). The $\delta^{18}O$ record of the surface-dwelling planktonic formaminifer *G. ruber* (white) has recorded the monsoon-induced freshwater outflow from the Ganges-Brahmaputra ($-8‰$ $\delta^{18}O$) and related salinity changes. Salinity changes are calculated from alkenone temperature data (Sonzogni et al., 1998) and $\delta^{18}O$ data (Rostek et al., 1993) corrected by the global $\delta^{18}O$ ice effect (Vogelsang, 1990). Because the D-O $\delta^{18}O$ variations of the surface water of the Arabian Sea (Schulz et al., 1998) were smaller than 0.3‰, we assume that the average $\delta^{18}O$ value of the Ganges-Brahmaputra water was only slightly affected by variations in the source area of the Arabian Sea. Furthermore, modeling indicates that even the great climatic changes from glacial to interglacial periods had minimal effects on the $\delta^{18}O$ value of the precipitation in the drainage area of the two rivers (Jouzel et al., 2000).

INTENSITY CHANGES OF THE MONSOON IN ASIA

The variability of the $\delta^{18}O$ record of the surface water is thought to be caused by variations of the intensity of the summer monsoon. An estimation of the paleosalinities based on the $U^{k'}_{37}$ temperatures supports this interpretation (Fig. 3). Before the last glacial maximum (LGM), temperatures were slightly lower than modern ones and decreased during the LGM by only 2 °C. Because of the small temperature changes, the $\delta^{18}O$ record is a good approximation of the fluvial input or the summer monsoon intensity. Between 80 and 30 ka, the

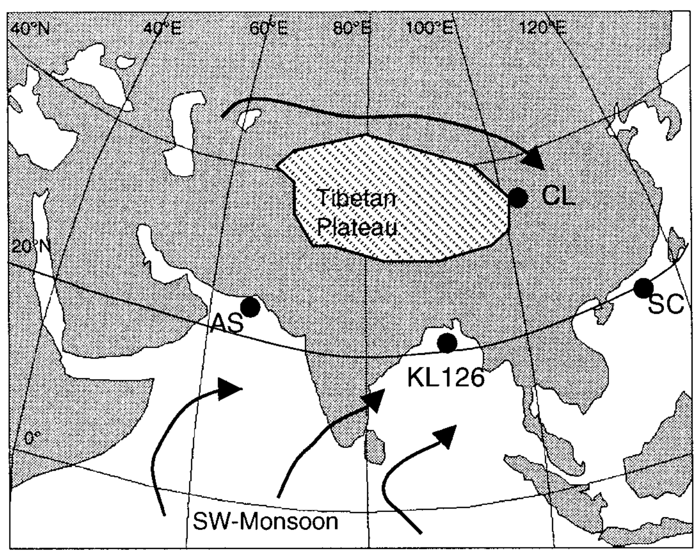

Figure 1. Flow of humid, warm air during summer monsoon across Arabian Sea and Bay of Bengal (shown by arrows) to central southeastern Asia and records used for paleoclimatic reconstruction (shown by dots; KL 126—Bay of Bengal [this study], SC—South China Sea [Wang et al., 1999], AS—Arabian Sea [Schulz et al., 1998], and CL—Chinese loess area [An Zhisheng, 2000]).

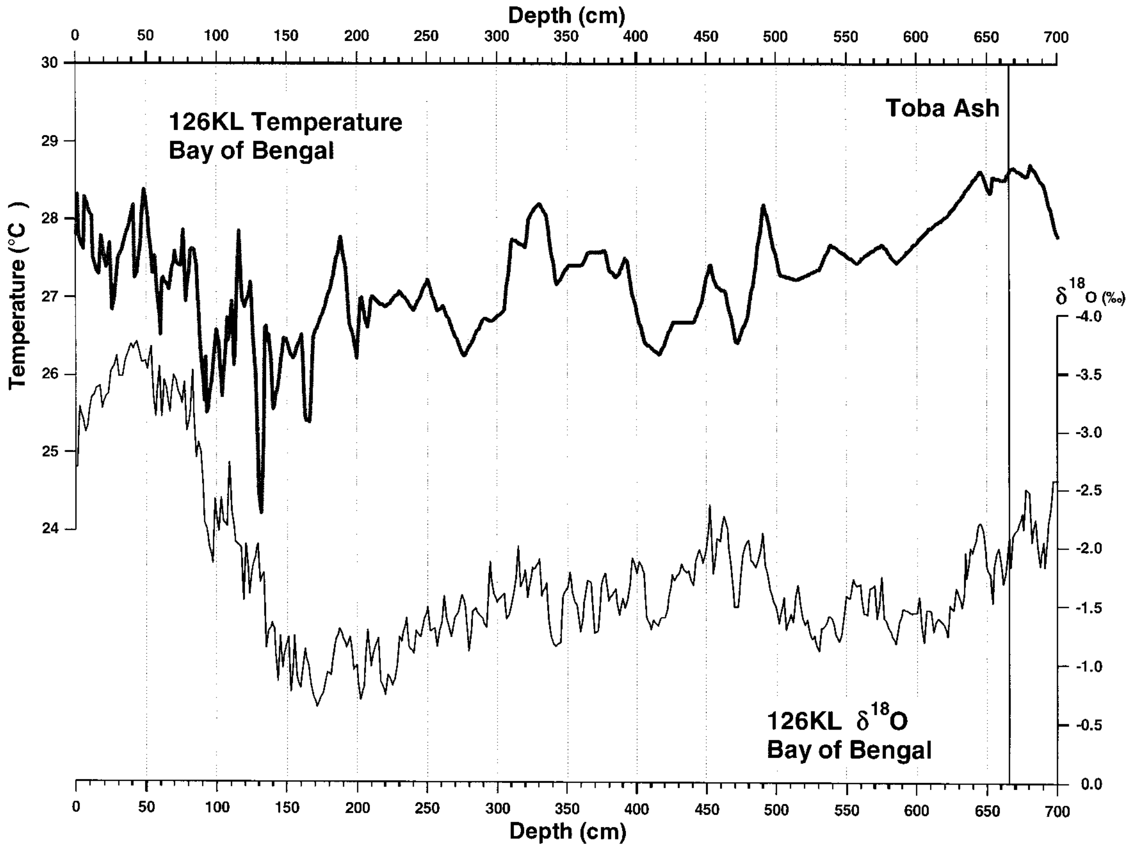

Figure 2. Variations of marine δ^{18}O stable isotope record of planktonic foraminifera *Globerigerinoides ruber* and surface temperature based on U^{K}_{37} index (Sonzogni et al., 1998) versus depth in core KL 126 from hemipelagic sediments in northern Bay of Bengal. Thick line marks position of volcanic ash of Toba megaeruption.

δ^{18}O values and consequently the summer monsoon intensity oscillated at intermediate levels between the LGM and Holocene values. The slow long-term increase was modified by D-O-like variations; a rapid decrease of as much as 0.75‰ was followed by a gradual slow increase. D-O events 21, 20, 19, and 14 are especially well documented. The LGM is characterized by the highest δ^{18}O values, and during that period the δ^{18}O gradient of *G. ruber* (white) tests in the Bay of Bengal declined from the recent 1.5‰ to 0.5‰ (Duplessy, 1982). The high values, the low gradient, and the composition of planktonic foraminifera (Cullen, 1981) indicate that the freshwater flow was strongly reduced in that period. In addition, the minimal δ^{18}O difference between the surface water of the Arabian Sea (Schulz et al., 1998) and the Bay of Bengal also confirm that the southwestward transfer of water from the Arabian Sea to India and the Himalayas nearly stopped (Duplessy, 1982). During the transition from the LGM to the Holocene, δ^{18}O values rapidly oscillated due to coupled global and local processes, e.g., influx of meltwater into the oceans from the polar regions and the Himalayan snow cover and the stepwise reestablishment of the summer monsoon (Gasse and Van Campo, 1994).

The rapid decreases of δ^{18}O values starting at 14 200 and 11 800 yr B.P. correlate with the global meltwater spikes MWP1A and MWP1B and the associated rise in sea level (Fairbanks, 1989). The low $U^{k\prime}_{37}$ temperatures and the shifts by 1.0‰ and 1.5‰ indicate an addition of cold, oxygen isotope-light meltwater from the Himalayas. The Younger Dryas exhibits the usual relapse to conditions similar to LGM. In the middle Holocene δ^{18}O values reached a minimum due to increased summer monsoon intensity (Feng et al., 1999; Sirocko et al., 1996).

The interpretation of δ^{18}O variation in terms of changing summer monsoon intensity is supported and supplemented by paleoclimatic records from other monsoon areas. Multiple facies changes in the Arabian Sea, especially that caused by the paleoproductivity of surface water, are associated with the D-O events (Schulz et al., 1998). The Toba ash confirms the synchronous development on both sides of the Indian subcontinent.

The summer monsoon–induced fluvial runoff into the South China Sea is perfectly imaged in a high-resolution 38 k.y. record from the upper continental slope southeast of the Pearl River mouth (Wang et al., 1999). The salinity changes determined from δ^{18}O of *G. ruber* (white) and the $U^{k\prime}_{37}$ temperature are generally in the same direction as the changes in the δ^{18}O record of the GISP2 core and conform with the record of the Bay of Bengal salinity decrease caused by increased summer monsoons during the D-O interstadials 1 to 20.

Cycles of aridity and humidity are well documented in the western Chinese loess and paleosol sequences, which can be correlated with D-O cycles dated by ^{14}C and infrared stimulated luminescence (Chen et al., 1997; Porter and An, 1995; Fang et al., 1999). The paleosols of the interstadials confirm the transport of moisture via the summer monsoon, whereas deposition of coarse dust during stadials indicates dry conditions with heavy winds. The rate of dust deposition increased considerably during the cold stadials and the LGM (Overpeck et al., 1996a).

High-resolution records from the most sensitive areas of the Indian and East Asian monsoon—the Bay of Bengal, northern Arabian Sea, South China Sea, and western Chinese loess plateau—prove that the paleoclimatic history reflects the summer monsoon variability and especially its capacity to transfer moisture from the Indian Ocean into eastern Asia. Results of large-scale climate modeling (Barnett et al., 1988; Meehl, 1994) and observations of recent variability (Yang, 1996) indicate that enhanced winter and spring snow cover on the Tibetan Plateau delays the onset and reduces the intensity of the following

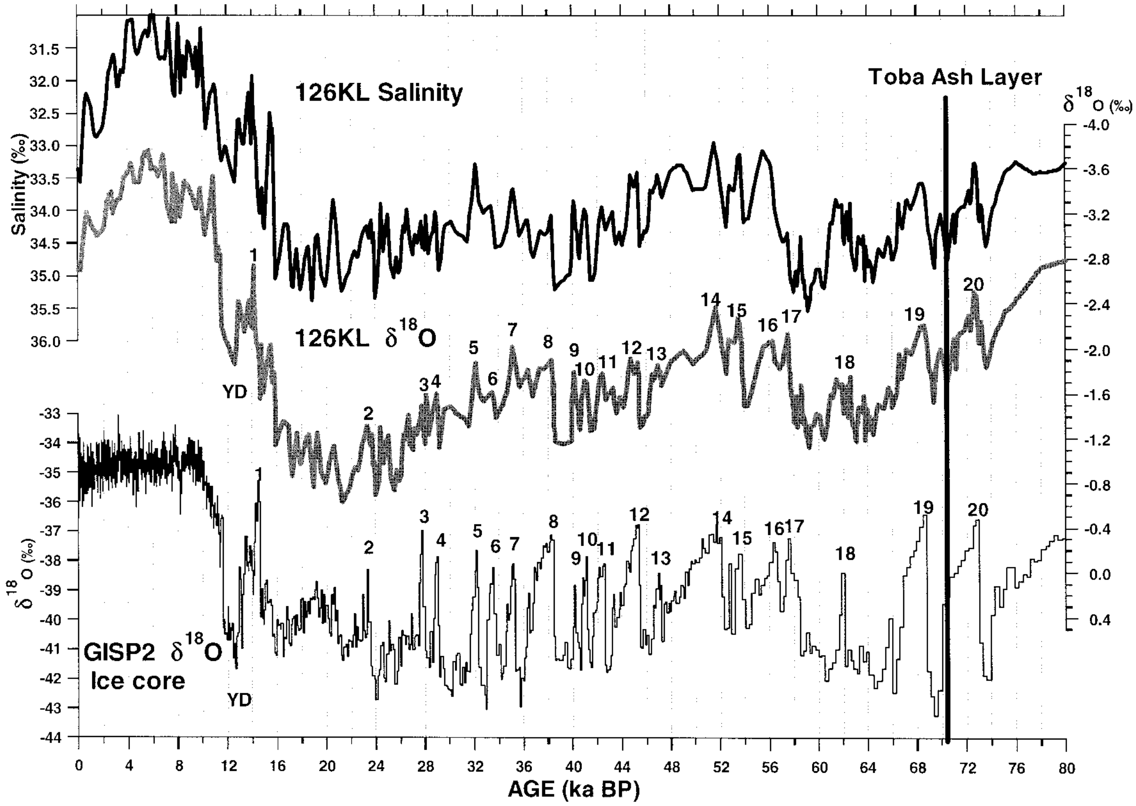

Figure 3. Correlation of salinity changes and marine $\delta^{18}O$ values in northern Bay of Bengal with Greenland GISP 2 $\delta^{18}O$ ice record (Dansgaard et al., 1993). Salinity is calculated from marine $\delta^{18}O$ and $U^{k'}_{37}$ temperatures based on data shown in Figure 2. Numbers indicate Greenland interstadial IS 1–20; YD is Younger Dryas. Layer of volcanic ash from Toba megaeruption, and its traces in Greenland ice core (Zielinski et al., 1996), is used as independent time marker for correlation.

summer monsoon (Barnett et al., 1988). Our study of the paleoclimatic records demonstrates that this effect can also explain the variability of summer monsoon in the past.

IMPLICATIONS OF THE MONSOON SYSTEM FOR THE NORTHERN HEMISPHERE

Within the precision of the dating techniques used, the summer monsoon variations are synchronous with the D-O cycles of the Greenland ice cores (Schulz et al., 1998; Wang et al., 1999; An Zhisheng, 2000). Many of the D-O and summer monsoon cycles, especially between 80 and 45 ka, start with a rapid shift followed by a gradual change. At the height of the wet (central Asia), warm (Greenland ice) interstadial conditions, climate gradually changes to drier (central Asia), colder (Greenland), and dustier (central Asia, Greenland) stadial conditions. Cold, dry conditions prevail for a short period and the cycle is terminated by a rapid shift to warm, wet conditions.

D-O cycles are also known in many high-resolution marine and continental climatic records from the Northern Hemisphere (review by Alley et al., 2000). The D-O cycles of the North Atlantic, in particular, are thought to be caused by variations of freshwater input or iceberg discharges that influence the thermohaline circulation (Broecker, 1994) and to result from instability of the Greenland ice sheet (Bond and Lotti, 1995; Kreveld et al., 2000). Variations in the oxygen minimum zone in the Santa Barbara basin off the southern California coast are synchronous with the D-O cycles and have been interpreted to be due to atmospheric warming by greenhouse gases (Hendy and Kennett, 1999). The wide occurrence and the apparent synchronous appearance of various D-O type climatic variations in different oceans suggest that the cycles might be caused by atmospheric circulation and ocean ice cover (Mayewski et al., 1994), atmospheric teleconnections (Hostetler et al., 1999), dust aerosols (Overpeck, 1996b), tropical heat budgets (McIntyre and Molfino, 1996; Cane, 1998), or equatorial wind stress (Klinck and Smith, 1993).

On the basis of records from India, Southeast Asia, China, and adjacent seas, we assume that these cycles in the Northern Hemisphere are caused by coupled processes of snow and dust accumulation during the Asian monsoons. The summer monsoon reaches its greatest strength at the warmest time of an interstadial, increasing the humidity in central Asia. Part of the moisture, however, accumulates over the years as snow at the rim of the Tibetan Plateau, and the increasing humidity and the higher albedo of the expanding snow cover in central Asia gradually reduce the summer pressure gradient between the warm continent and the cold Indian Ocean and, hence, the intensity of the summer monsoon (Overpeck, 1996a; Sirocko et al., 1996). With increasing snow accumulation and lower snow line over several hundred years, summer monsoon precipitation gradually diminishes and migrates southward. In the following stadial period, the moisture-transporting monsoon is replaced by dry westerly winds, which transport dust north and south of the Tibetan Plateau to the snow-covered areas (An Zhisheng, 2000). With the prevailing dry conditions, some of the snow slowly disappears by sublimation. Dust settling on the snow decreases the albedo and, in combination with the warming by the worldwide peak of dust aerosols (Overpeck et al., 1996b), triggers the massive and rapid return of the summer monsoon, introducing the next interstadial.

The main effect of the summer monsoon is the large variation in the amount of water vapor injected into the atmosphere over large parts of eastern Asia, profoundly influencing the conditions that produce the greenhouse effect (Raval and Ramanathan, 1989). Some of this water vapor may be widely distributed in the Northern Hemisphere by the westerlies, like the dust found in the Greenland ice originating in the desert areas of eastern Asia (Biscaye et al., 1997). The changes in the marine and continental distribution of water vapor with the D-O cycles of the summer monsoon also affected the snow accumulation rates on the ice sheets of the Northern Hemisphere, especially in the period of advancing ice sheets and falling sea level, when generally more water was deposited as ice onshore than flowed back into the ocean. During an interstadial, high rates of transport of water vapor from marine areas to continental areas increased continental snow and ice accumulation (Kapsner et al., 1995) and caused a falling sea level, while during the dry and dusty stadial a more balanced ocean-continent budget and a stable sea level can be expected. Because the seaward flow of ice is more continuous than the accumulation rates, the sea level might have even risen, especially near the end of an extended stadial. Unfortunately, the currently available data on sea-level changes do not have the resolution needed to monitor variations at D-O time scales (Chappell et al., 1996). This postulated probably small rise of sea level during a stadial may have caused the release of the pulses of freshwater and icebergs of the Heinrich events, affecting the thermohaline circulation in the North Atlantic. Thus, this conceptual model may explain how changes in the intensity of the summer monsoons have influenced the development of the Northern Hemisphere climate.

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