Century-scale events in monsoonal climate over the past 24,000 years

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BOTH the marine sediment record and numerical modelling of the atmospheric summer circulation over the northern Indian Ocean and southeast Asia have shown that the monsoonal climate exhibits a direct but nonlinear response to the intensity of solar insolation during summer, with a time lag of several thousand years[1,2]. Here we present evidence from a high-resolution record of oxygen isotopes and carbonate spanning the past 24,000 calendar years that the response of the southwest monsoon over the Arabian Sea to long-term, gradual insolation changes occurred in several distinct events of less than 300 years duration, at 14,300, 13,500, 13,060, 9,900, 8,800 and 7,300 ^{14}C yr BP. Thus, during this transitional period from glacial to post-glacial conditions the slow solar forcing seems to have induced very rapid changes in local climate. We speculate that the rapid response may be related to albedo changes in Asia.

Our objective was to monitor the chronology of the evolution of the monsoonal system over southeast Asia since the last glaciation. The Arabian Sea is especially suited to this because its deep-sea sediments permit extraction of continuous records of wind-driven oceanic upwelling, continental humidity, and dust and river discharge.

We sampled core 74KL from the upwelling region of the western Arabian Sea (Fig. 1) at 2.5-cm intervals, which corresponds to an average time resolution of 300 calendar years (Table 1). The core was obtained from the East Sheba Ridge 1,000 m above the turbidite plain of the Indus fan, 300 km south of the Arabian continental margin. Sediment structures suggest that the record of the uppermost 2 m of core 74KL is continuous and undisturbed. The stratigraphic base of 74KL is the δ^{18}O record of the surface-dwelling planktonic foraminifera *Globigerinoides ruber*. Its oxygen isotope composition decreased from glacial values near 0‰ to Holocene values averaging near 1.75‰ (Fig. 2), punctuated by a series of abrupt changes. Of this range, 1.2‰ can be explained by changes in the global ice volume[3,4], but 0.55‰ should be of local origin. This local contribution must be reflected in the abrupt events, because these are of far larger amplitude than can be accounted for on the basis of changes in the global δ^{18}O value of sea water owing to ice-sheet melting. For example, the 0.5‰ change during event 5 is larger by 0.35‰ than the global increase of 0.15‰ owing to the meltwater pulse Ib that occurred at this time[5].

To verify that the abrupt changes observed in core 74KL are not confined only to the upwelling areas off Arabia, we compared them with high-resolution records of cores KS8 and 422 from the Gulf of Oman, 900 km further north (Fig. 1). These also show the δ^{18}O events 2, 3, and 5, and a decrease in both grain sizes and wind-borne lithogenic carbonate content during event 3.

These abrupt shifts towards lighter δ^{18}O values in all three cores can be related to four possible mechanisms: (1) increased precipitation or decreased evaporation, leading to lower salinities and lower δ^{18}O values in the Arabian Sea surface water[6]; (2) an increased proportion of isotopically light Persian Gulf outflow water (δ^{18}O: 0.7‰ SMOW at 200 m depth) and Red Sea outflow water (δ^{18}O: 0.25–0.4‰ SMOW, at 200–1,200 m depth in the outer Gulf of Aden) that are admixed to the isotopically enriched South Indian Central Water (δ^{18}O: 1.3‰ SMOW, 100–200 m depth) and Arabian Sea surface waters (δ^{18}O: 0.9‰ SMOW at 0–100 m depth)[7-9]; (3) a change in the seasonality of *G. ruber* growth, which today reveals two seasonal maxima of production, one with more positive values during the nutrient-rich summer months and a second one with more negative values in the warm winter surface waters[10,11]; (4) an increase in sea surface temperature (SST). This change, however, seems unlikely, because it would imply warmer SST during the Holocene, when upwelling was increased and SSTs were colder than during the glacial period[12].

Which of these possible explanations is the actual cause for the abrupt steps towards lighter δ^{18}O values cannot be resolved at this stage. However, in every case, these changes occurred within less than 300 years (in fact, within less than 120 years in core 422; FS manuscript in preparation) and abrupt transitions are found at 13,500; 13,060; 9,900; 8,800; and 7,300 ^{14}C yr BP (Fig. 2). The δ^{18}O decrease at 12,000 ^{14}C yr BP is only a spike and does not qualify for an event, which we define as a change from one sample to another, reaching a new level of similar values. All of the abrupt steps in the δ^{18}O record can be regarded as

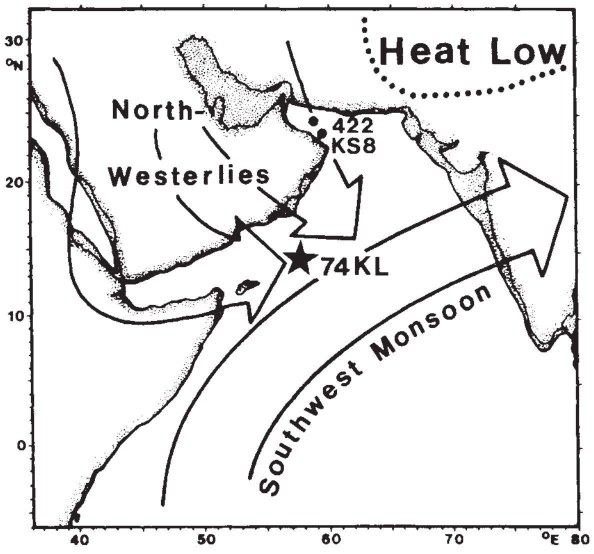

FIG. 1 The Arabian Sea during summer season. A thermal heat low over Asia attracts rain-bearing southwesterly winds from the Southern Hemisphere and dry northwesterlies from Arabia, the northern branch of which carries huge amounts of dolomite (up to 60% of total dust load in aerosol samples in Kuwait)[34]. Stippled areas indicate dry land during glacial times of lowered sea level. Sediment cores: 74KL (14° 19.26′ N, 57° 20.82′ E, 3,212 m); KS8 (23° 28.00′ N, 59° 11.50′ E, 2,900 m): 422 (24° 23.40′ N, 59° 02.50′ E, 2,732 m).

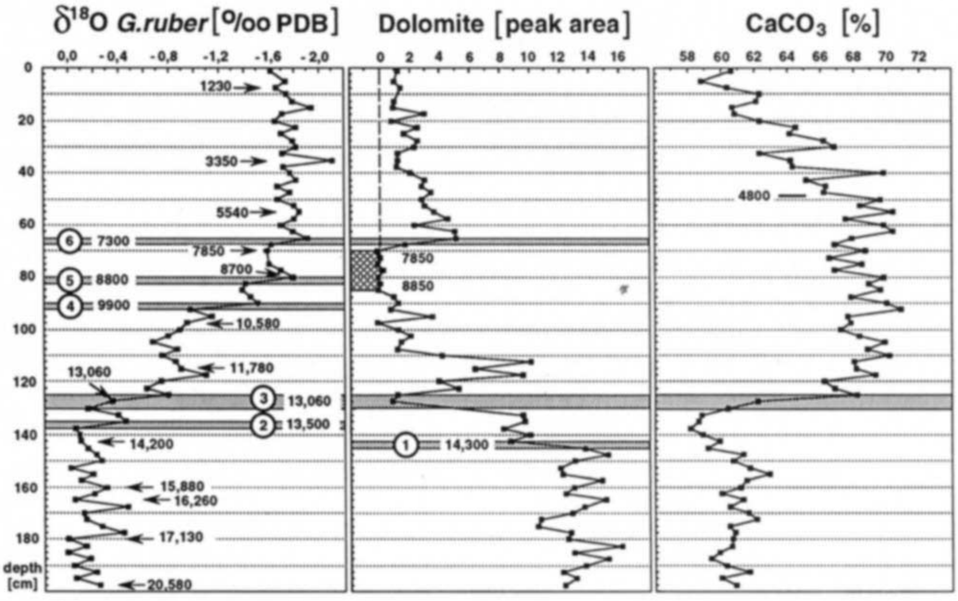

FIG. 2 AMS radiocarbon ages of core 74 KL (Table 1) and $\delta^{18}O$ values of *G. ruber*, diameter: 315–400 μm, 20 specimens, cleaned in methanol and ultrasonic bath, measured at the ^{14}C laboratory of Kiel University, Dolomite was measured with a Philips diffractometer at 2,89 Å in the lithogenic fraction› 2 μm. Values represent the peak area from 35.7° to 36.5° 2θ (Co Kα) and can be regarded as being comparable to percentage (compare percentages in refs [13, 14]). The sharp increase in $CaCO_3$ content at 130–125 cm is most probably caused by a reduction of lithic components, the flux of which we quantified as 43 g m^{-2} yr^{-1} during the last glacial and 18 g m^{-2} yr^{-1} during the Early Holocene and 29 g m^{-2} yr^{-1} during the Late Holocene (F.S., manuscript in preparation). The hatched area between 85 cm and 70 cm depth indicate the depth interval of the Early Holocene southwest monsoon maximum, whereas the short line at 50 cm depth indicates the knickpoint (change in profile) towards the Late Holocene aridification.

being of mostly local origin, because the glacial/interglacial contrast in distribution patterns of $\delta^{18}O$ (ref. 6) had revealed large local amplitudes, and there are no other records worldwide to show the same succession of events.

Beside the hydrographic events, we observe changes of similar abruptness in the lithic sediment supply, which is exclusively of aeolian origin in the western Arabian Sea[2], stemming from dust plumes raised in Arabia and Mesopotamia during spring and summer, and transported to the Arabian Sea by northwesterly winds (Fig. 1)[13,14]. The provenance of the lithic aerosol fraction in these dust plumes is best recorded by the occurrence of dolomite, which is preferentially transported in the northern branch of the northwesterlies, because dolomite is derived from Mesozoic strata that are prominent only in northern Arabia, or from sebkha sediments around the Persian Gulf, characterized by the frequent occurrence of protodolomite. The $CaCO_3$ content of the Arabian Sea sediments represents another record of lithic sediment supply, because the carbonate percentages in the open ocean at less than 3,500 m water
depth are largely controlled by dilution with lithic matter[2,14].

The northwesterly winds that carry these dust loads and also the southwest monsoon are both attracted by the Asian heat low in such a way that the southward (eastward) extent of the

TABLE 1 ^{14}C AMS ages

AMS sample depth (cm)	AMS ^{14}C age (^{14}C yr BP)	Error 1-sigma (yr)	Calibrated age (cal. yr BP)	Reference for calibration	Event	Event depth (cm)	Calibrated event age (cal. yr BP)	Interpolated event age (^{14}C yr BP)
7.50	1.230	60	1,170	30				
35.00	3,350	100	3,632	31				
					Aridification	48.75	5,500	4,800
55.00	5,540	90	6,320	31				
					Event 6:	66.25	8,050	7,300
70.00	7,850	90	8,625	32	Humid interval	70.00	8,600	7,850
80.00	8,700	120	9,550	33				
					Event 5:	81.25	9,700	8,800
					Humid interval	82.50	9,900	8,850
					Event 4:	91.25	11,450	9,900
97.50	10,580	120	11,880	33				
115.00	11,780	130	13,780	28				
127.50	13,060	150	16,060	28	Event 3:	127.50	16,060	13,060
					Event 2:	136.25	17,000	13,500
142.50	14,200	170	17,700	28				
					Event 1:	143.75	17,800	14,300
160.00	15,880	170	19,380	28				
165.00	16,260	170	19,760	28				
180.00	17,130	180	20,630	28				
197.50	20,580	260	24,080	28				

Dating: ^{14}C AMS ages of core 74KL, analysed on 10 mg of *G. ruber*, diameter 315–400 μm, at the Centre des Faibles Radioactivités, CNRS, Gif sur Yvette, France. A 400-yr correction is applied for the age of the sea water. Direct measurements [23] of the age of sea water in the coastal upwelling water of Oman revealed ages of 800 yr, so our use of a 400-yr correction has some uncertainties. Carbon-14 ages for events were interpolated from the U/Th-corrected ^{14}C AMS age profile and then reconverted to the conventional ^{14}C scale[28,29] using the switch points at 9,100 ^{14}C yr = 9,800 calendar yr BP; 10,400 ^{14}C yr = 12,400 calendar yr BP; and 13,100 ^{14}C yr = 15,100 calendar yr BP, 13,200 ^{14}C yr = 16,700 calendar yr BP.

northwesterlies outlines the northernmost (westward) extent of the southwest monsoon winds[15] (Fig. 1). High dolomite flux thus indicates a weak southwest monsoon and low dolomite indicates a strong southwest monsoon. To complicate the picture, on glacial/interglacial scales the dolomite supply must also be affected by sea-level changes, because the Persian Gulf was dry land during glacial times. It was most probably a dust-entrainment area, which was lost during the transgression into the Gulf[14,16].

The initial onset of climate change in Southeast Asia at the end of the glacial is marked by event 1, 14,300 ^{14}C yr BP. This decrease of dolomite content must have been due to changes in the atmospheric circulation, because sea level was still 110 m below the present level and therefore well below the transgression level into the Gulf (100–65 m below modern surface waters)[16]. During event 3 (13,060 ^{14}C yr BP), which represents the major transition between the glacial and the Holocene, however, we observe a simultaneous sharp change of δ^{18}O, dolomite and $CaCO_3$ content.

A brief climatic extreme is recorded by the total absence of dolomite dust input from 8,850–7,850 ^{14}C yr BP, when dust plumes from the Persian Gulf region did not reach as far south as our core position. The continental heat low was strongest during this time[1] and the track of the southwest monsoon and its associated pattern of precipitation can be expected to have shifted to their northernmost position. Indeed, there was widespread moisture all over East Africa, Arabia, Pakistan, northern India, and even Tibet, characterizing an early Holocene humid interval[17–21]. This indicates that the convergence between the southwest monsoon and the northwesterlies, which occurs mostly over the ocean today, was then located over inner Arabia. Judging from the actual age of the humid interval (9,900–8,600 calendar yr BP, Table 1), it followed immediately after the maximum summer insolation at 20° N to 35° N, 10,000 calendar years ago[22].

The middle Holocene was then characterized by the return of dolomite after 7,300 ^{14}C yr BP (event 6) and a general increase in dust flux (decrease in % $CaCO_3$) after 4,800 ^{14}C yr BP, equal to 5,500 calendar yr BP, which indicates a middle Holocene increase in Arabian aridity. This deterioration is reflected all over North Africa and Arabia by lowered lake levels[17,20,21], a decline of vegetation cover[18,19] and, especially, the formation of hierarchical societies in the then overpopulated Nile valley and Mesopotamia starting near 5,300 calendar yr BP. In contrast, neolithic settlements in the inner desert of Arabia have been completely abandoned between 5,600 and 5,200 calendar yr BP[23].

To infer the mechanisms that caused these abrupt transitions in the monsoonal system, we focus on the largest event of atmospheric and hydrographic change, event 3 (13,060 ^{14}C yr BP). This was at the very beginning of global meltwater pulse Ia, which started at 13,000 ^{14}C yr BP and culminated about 12,300 ^{14}C yr BP ago[5]. One possible mechanism that could explain such an early monsoon response is a reaction to albedo changes in Asia during the initial phases of deglaciation, when areas of snow fields at high elevation disappear and expose the soils to the slowly increasing solar radiation. This mechanism would be fast enough to trigger abrupt climate changes. Its effect would be a sharp increase in the length of the summer season, with effects on the seasonal patterns of wind trajectories as well as on the strength and duration of the upwelling season.

Sensitivity experiments with global circulation models have indeed shown that albedo changes induced by changing snow cover exert considerable control over the development of the continental heat low over mainland Asia, and affect the strength and duration of the southwest monsoon winds over the Arabian Sea[1,24,25]. Accordingly, a synchronous evolution of deglaciation events and southwest monsoon intensification should be physically related. Modern meteorological data reveal that in general a cold winter over the Northern Hemisphere is followed by a weakened heat low and weakened southwest monsoon intensity during the next summer[26].

Time-series work on monsoonal change over the past 350,000 years demonstrates that the response of atmospheric circulation, coastal upwelling, and continental aridity lagged the changes in solar forcing by several thousand years[2]. The succession of atmospheric and hydrographic events in the Arabian Sea shows, however, that these transitions are complicated and are observed not as a gradual response in each part of the climate system, but as a number of very brief evolutionary steps, interdependent in the atmosphere and the ocean.

There might, however, be a stable time-related pattern in the succession of events. When converted to a calendar year timescale (Table 1), the interval between events 5–6, 4–5 and 1–3 lasted about 1,700 years. This recurrence period may be characteristic of the internal variability in the monsoon region; a similar period was found in the δ^{13}C record of *G. ruber* in core 74KL, and is also known from SST variations in the East Atlantic upwelling belt[27]. □

Received 22 October 1992; accepted 19 May 1993.

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ACKNOWLEDGEMENTS. The study was funded by the German Program of Climatic Research (Bundesministerium für Forschung und Technik). We thank K. Scheffer for technical assistance and W. Prell for constructive suggestions. The writing was in part done when F.S. held a fellowship of the Max Kade Foundation at Lamont Doherty Earth Observatory. J.C.D. and M.A. are supported by CEA, CNRS and INSU (PNEDC).