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The southwest Indian Monsoon over the last 18000 years

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Abstract. Previously published results suggest that the strength of the SW Indian Monsoon can vary significantly on century- to millenium time scales, an observation that has important implications for assessments of future climate and hydrologic change over densely populated portions of Asia. We present new, well-dated, multi-proxy records of past monsoon variation from three separate Arabian Sea sediment cores that span the last glacial maximum to late-Holocene. To a large extent, these records confirm earlier published suggestions that the monsoon strengthened in a series of abrupt events over the last deglaciation. However, our data provide a somewhat refined picture of when these events took place, and suggest the primacy of two abrupt increases in monsoon intensity, one between 13 and 12.5 ka, and the other between 10 and 9.5 ka. This conclusion is supported by the comparisons between our new marine data and published paleoclimatic records throughout the African-Asian monsoon region. The comparison of data sets further supports the assertion that maximum monsoon intensity lagged peak insolation forcing by about 3000 years, and extended from about 9.5 to 5.5 ka. The episodes of rapid monsoon intensification coincided with major shifts in North Atlantic-European surface temperatures and ice-sheet extent. This coincidence, coupled with new climate model experiments, suggests that the large land-sea thermal gradient needed to drive strong monsoons developed only after glacial conditions upstream of, and on, the Tibetan Plateau receded (cold North Atlantic sea-surface temperatures, European ice-sheets, and extensive Asian snow cover). It is likely that abrupt changes in seasonal soil hydrology were as important to past monsoon forcing as were abrupt snow-related changes in regional albedo. Our analysis suggests that the monsoon responded more linearly to insolation forcing after the disappearance of glacial boundary conditions, decreasing gradually after about 6 ka. Our data also support the possibility that significant century-scale decreases in monsoon intensity took place during the early to mid-Holocene period of enhanced monsoon strength, further highlighting the need to understand paleomonsoon dynamics before accurate assessments of future monsoon strength can be made.

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Introduction

The southwest Indian Monsoon system is one of the major climate systems of the world, impacting large portions of both Africa and Asia. The modern large-scale time-averaged seasonal variations in this monsoonal system are fairly well understood, and are closely linked to the greater heat capacity of the ocean relative to the surrounding land masses (Hastenrath 1991; Fein and Stephens 1987). In the Northern Hemisphere summer, the Tibetan Plateau warms rapidly relative to the Indian Ocean. The resulting low pressure over Asia and higher pressure over the ocean gives rise to the strong low-level atmospheric pressure gradient that in turn generates the SW monsoon. In years of low snowfall, the Tibetan Plateau is able to warm earlier and generate a stronger monsoonal circulation (Dickson 1984; Barnett et al. 1988, 1989; Meehl 1994). Deep snow depths, and associated influences on albedo and soil hydrology, delay and weaken the monsoon (Barnett et al. 1989). In the winter, the continent cools relative to the ocean, the pressure gradient completely reverses, and the dominant flow across the Arabian Sea becomes northeasterly.

Terrestrial and marine paleoclimatic records indicate that century- to millennia-scale variations in the monsoon are larger than any observed over the past century, and that these variations are likely to have multiple causes (Kutzbach 1981, 1987; Street-Perrott and Roberts 1983; Street-Perrott and Perrott 1990; Clemens et al. 1991; Prell and Kutzbach 1992; Ander-

Table 1. List of sites with paleomonsoon records shown in Figs. 1 and 2

Site number	Site or core name	Reference(s)
1	Oyo	Ritchie et al. 1985; Ritchie 1994
2	Bir Atrun (El' Atrun Oasis)	Ritchie 1987; Ritchie and Haynes 1987
3	Lake Sidigh complex	Pachur and Hoelzmann 1991
4	Selima Oasis	Ritchie and Haynes 1987
5	Wadi Howar	Pachur and Kröpelin 1987; Kröpelin and Soulie-Marshe 1991
6	Muchoya Swamp	Taylor 1990
7	Ahakagyezi Swamp	Taylor 1990
8	Lake Kivu	Haberyan and Hecky 1987
9	Kuruyange	Bonnefille et al. 1991
10	Lake Cheshi	Stager 1988
11	Lake Tanganyka	Gasse et al. 1989; Haberyan and Hecky 1987; Vincens 1989, 1991
12	Lake Rukwa	Haberyan 1978
13	Lake Victoria	Adamson et al. 1980
14	Mt. Satima mire	Street-Perrott and Perrott 1990
15	Lake Turkana	Owen et al. 1982
16	Lake Ziway-Shala complex	Gasse and Street 1978; Street-Perrott and Perrott 1990
17	Lake Abhe	Gasse and Street 1978; Gillespie et al. 1983
18	Lake Afrera	Gasse and Street 1978; Gillespie et al. 1983
19	Lake Asal	Gasse and Street 1978; Gillespie et al. 1983
20	Rub' al Khali	McClure 1976
21	Nizwa complex	Clark and Fontes 1990
22	Marine core RC27-23	New data – this paper
22	Marine core RC27-28	New data – this paper
22	Marine core RC27-14	New data – this paper
23	Lunkaransar	Bryson and Swain 1981; Swain et al. 1983
24	Didwana Lake	Bryson and Swain 1981; Swain et al. 1983; Singh et al. 1990
25	Sumix Co	Gasse et al. 1991; Van Campo and Gasse 1993
26	Southern Tibet lakes	Fang 1991

son and Prell 1993). Terrestrial records of past change from Asia and Africa indicate that the monsoon was significantly weaker than present during glacial times (ca. 18 ka), much stronger than present during the early to mid-Holocene (ca. 9 to 5 ka), and weaker up to the present-day (Street-Perrott and Perrott 1990; Porter et al. 1992; An et al. 1993; Jarvis 1993; Winkler and Wang 1993; references cited in Table 1). Marine records of past change from the Arabian Sea yield a similar record of change (Duplessy 1982; Van Campo et al.1982; Prell 1984; Prell et al. 1990; Sirocko et al. 1991). Climate model simulations and available paleoclimatic data suggest that two mechanisms exert the dominant forcing on millennial-scale variations in monsoon strength. First, changes in the orbit of the Earth, predominantly in the precession of the equinoxes, control the amount of insolation reaching the Earth as a function of season, and hence the ability of the Tibetan Plateau to warm in the summer (Kutzbach 1981; Prell 1984; Kutzbach and Street-Perrott 1985; Prell and Kutzbach 1987; COHMAP 1988; Street-Perrott and Perrot 1990; Clemens and Prell 1990; Prell and Kutzbach 1992; Winkler and Wang 1993). Second, changes in glacial boundary conditions [ice volume, sea surface temperature (SST), albedo, and atmospheric trace-gas concentrations] have been cited as mechanisms that can alter the way in which the monsoon can respond to astronomical forcing (Duplessy 1982; Manabe and Broccoli 1985; Prell and Van Campo 1986; Prell and Kutzbach 1987; Overpeck et al. 1989; Street-Perrott and Perrott 1990; Gasse et al. 1991; Prell and Kutzbach 1992; Anderson and Prell 1993; Sirocko et al. 1993; deMenocal and Rind 1993).

Recently, the results of Clemens and co-workers (Clemens and Prell 1990, 1991; Clemens et al. 1991) have sparked an important controversy regarding the forcing of millennial-scale changes in monsoon strength. Based on the frequency-domain analysis of new 350-ky marine sediment records from two sites (the Owen Ridge, 300 km off the coast of Oman, and the southwestern Indian Ocean), they have argued that, while precession-forced insolation changes are the major pacemaker of monsoon strength, glacial boundary conditions have played only a minor role in determining the timing and strength of the Arabian Sea monsoon (Clemens and Prell 1991). More recently, several authors (Gasse et al. 1991; Prell and Kutzbach 1992; Anderson and Prell 1993; Sirocko et al. 1993; Van Campo and Gasse 1993) have confirmed that it is still important to recognize the central role of glacial boundary conditions in modifying the response of the monsoon to astronomical forcing. A purpose of our paper is to synthesize new well-dated 18-ky paleoclimatic time series from three monsoon-sensitive Arabian Sea sediment cores, along with available terrestrial data and model results, to produce a better understanding of how monsoon intensity was forced over the last precession cycle.

Recent results from ice-core and marine sediment records in the Greenland-North Atlantic region have

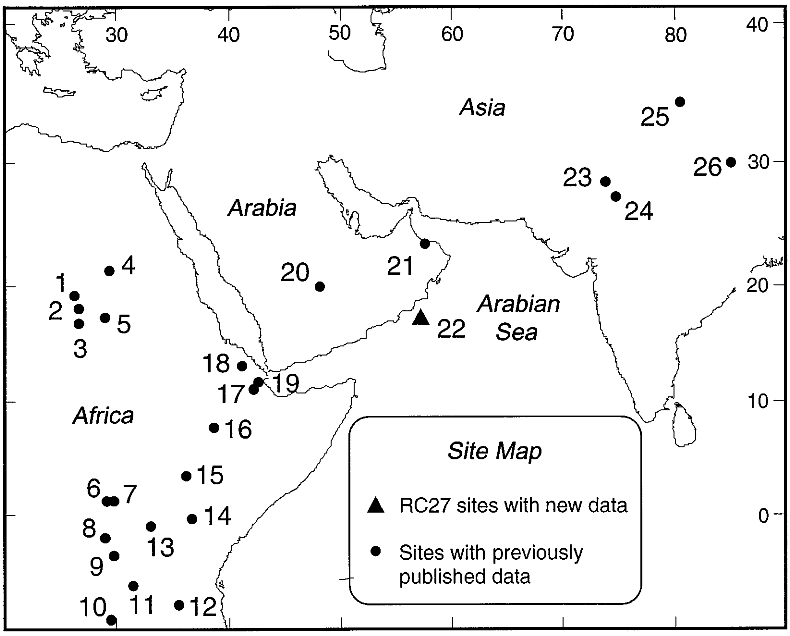

Fig. 1. Map of the Arabian Sea region showing location of sites discussed in the text. References for each site are given in Table 1

highlighted the capability of the climate system to shift abruptly between different climate states (Broecker and Denton 1989; Lehman and Keigwin 1992). Although much evidence points to salinity-driven changes in ocean thermohaline circulation as the primary forcing mechanism behind these abrupt glacial and deglacial changes (Manabe and Stouffer 1988; Stocker and Wright 1991; Paillard and Labeyrie 1994), synchronous abrupt climatic changes far from northern North Atlantic may point to alternative causal mechanisms. Our new data, analyses and model experiments, however, emphasize how abrupt changes in the North Atlantic may be invoked to explain abrupt changes at low latitudes and far from the North Atlantic.

Growing attention has also focused on the abrupt changes in monsoon strength that apparently took place during the Holocene, in the absence of significant changes in glacial boundary conditions (Street-Perrott and Perrott 1990; Gasse and Van Campo 1994). This is important to debates over how monsoon intensity will change in the future, because climate model results suggest that the future will be characterized by an enhanced monsoon (Houghton et al. 1992; Meehl and Washington 1993), just as the early to mid-Holocene was. Our new data support the existence of major century-scale changes in monsoon strength during the Holocene, but also points to the lack of a plausable theory to explain these changes. Additional data will be needed to define the exact spatio-temporal aspects of abrupt interglacial "warm-climate" climatic events, and to understand the processes that give rise to them.

New high-resolution records of monsoon variability

Marine sediments from the Oman margin of the Arabian Sea (Fig. 1) provide one of the best opportunities to study past variations in Indian Ocean monsoon variability because the sediments accumulate rapidly (>10 cm/ky), can be well dated and contain multiple proxy records of past monsoon variation. In this paper, we present new data from three cores collected approximately 150 km offshore Oman (RC27-14: 18.25°N, 57.66°E, 590 m water depth; RC27-23: 18.00°N, 57.59°E, 815 m; and RC27-28: 17.90°N, 57.59°E, 866 m; Fig. 1, Table 1). Each of the cores, selected from a suite of 26 collected during cruise RC2704, is characterized by homogenous green to olive green foraminifer-nannofossil oozes with $CaCO_3$ contents varying between 30–60%. We placed our records in a chronostratigraphic framework using $\delta^{18}O$ measurements and 35 AMS radiocarbon measurements on single species (>150 m size fraction) samples of planktonic foraminifera (Fig. 2). All of the AMS measurements were made at the Institut fur Mittelenergiephysik, Zurich and corrected by 400 year to account for the difference in radiocarbon content between surface waters and the atmosphere (Broecker et al. 1988). Only minor age differences were found between dates on each of seven pairs of coexisting species. With the exception of the lowest (below 270 cm) section of RC27-23, and a possible core hiatus near 145 cm in RC27-28, the core stratigraphies appear continuous and reliable. More complete descriptions and analyses of the cores will be published elsewhere.

Following the successful past use of foraminifera and pollen proxies for reconstructing paleomonsoon variability (Prell 1984; Prell et al. 1990; Van Campo et al. 1982; Prell and Van Campo 1986), we used these two independent proxies and standard methods (Prell 1984; Van Campo et al. 1982) to produce a multi-core multi-proxy record of monsoon variability spanning the past 20000 year. Pollen counts were only made for sections of cores with well-preserved pollen, and palynomorph identifications were made with the help of

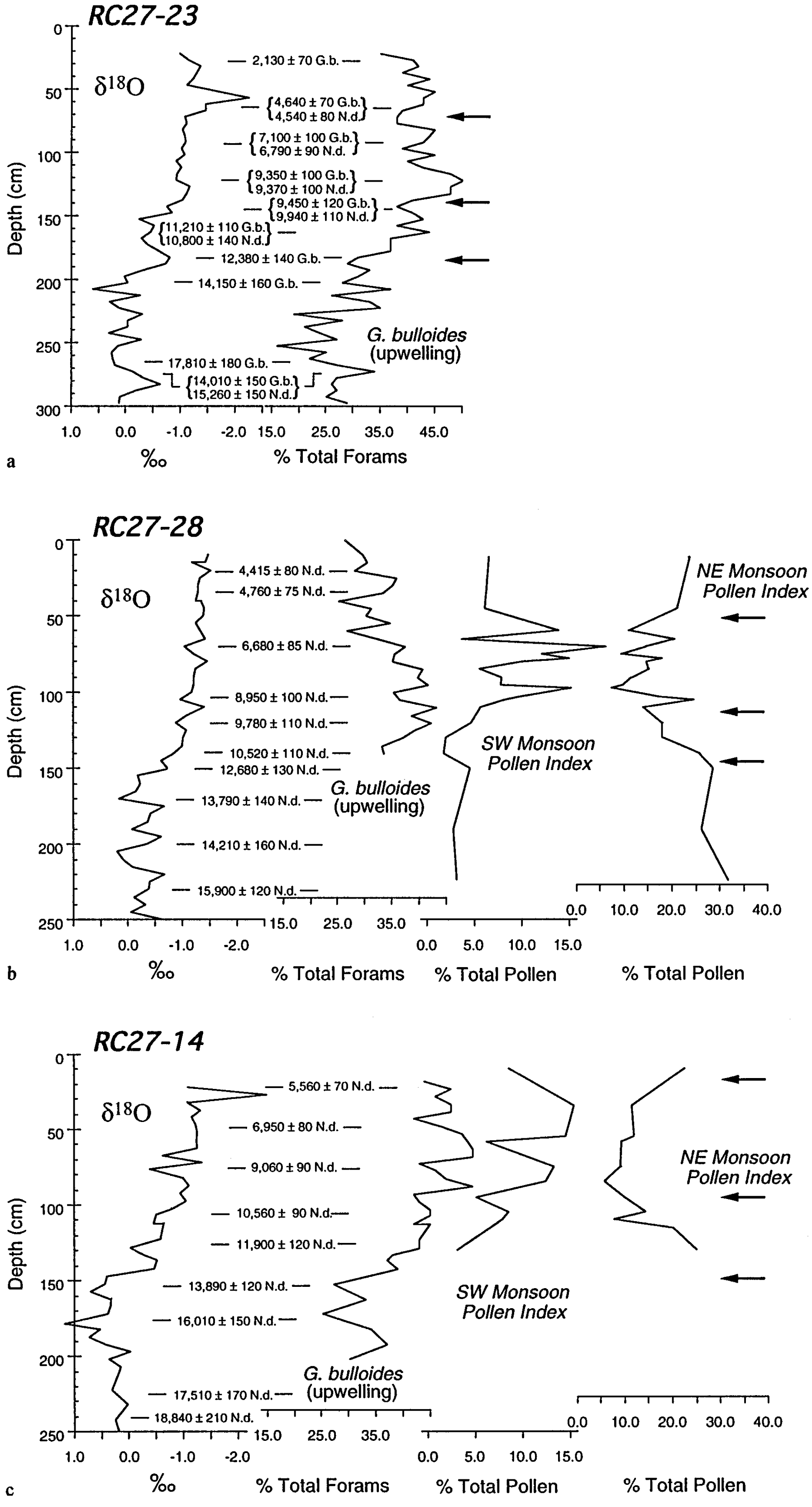

Fig. 2a–c. New paleoclimatic records from three Oman margin sediment cores. **a** RC27-23, **b** RC27-28, **c** RC27-14. In each case, oxygen isotope, percent *G. bulloides,* and percent pollen data are plotted where available. AMS radiocarbon dates are on single or paired single-species samples of planktonic foraminifera (G.b., *Globigerina bulloides;* N.d., *Neogloboquadrina dutertrei*). The SW monsoon pollen index was defined previously (Prell and Kutzbach 1987), and the NE monsoon pollen index is defined as the percentage of pollen indicative of Asian steppe in each sample (Table 2; Van Campo et al. 1982). Pollen data were only generated for core sections that had abundant and well-reserved fossil pollen, and are presented as a percentage of all fossil pollen found in a given sample. The *lower two arrows* in each plot mark the timing at which the SW monsoon increased in strength, whereas the *upper arrow* marks the approximate time the monsoon began to weaken from early Holocene maximum levels

the extensive pollen reference collection at the Laboratoire de Palynologie, Montpellier, France. Pioneering work with modern air sampling and sediments has illustrated the power of reconstructing monsoon variations using fossil pollen (Van Campo et al. 1982; Prell and Van Campo 1986; Prell and Kutzbach 1987). During the modern SW monsoon season (summer) and past periods of stronger SW monsoon flow, more poll-

Table 2. List of pollen taxa in the two monsoon pollen indices[a]

SW Monsoon pollen index	NE Monsoon pollen index
Avicennia	*Artemisia*
Combretaceae	*Calligonum*
Commiphora	*Centaurea*
Cyperaceae	Compositae lig
Indigofera	*Ephedra*
Kohautia	*Erodium*
Mallotus	Ombelliferae
Olea	*Plantago*
Phyllantus	
Podocarbus	
Rhizophoraceae	
Sapotaceae-Meliaceae spores	

[a] Following the convention of Van Campo et al. (1982), Prell and Van Campo (1986) and Prell and Kutzbach (1987). Data are presented (Fig. 2) as percentages of all fossil pollen found in a given sample

en typical of East African humid tropical vegetation typify shipboard air filters and sediment samples. In contrast, periods of weaker SW monsoon flow are characterized by larger proportions of pollen typical of the Asian steppe to north of the Arabian Sea. We follow the well-established convention (Van Campo et al. 1982; Prell and Van Campo 1986; Kutzbach and Prell 1987) of presenting the pollen data as two indices that discriminate between past strong southwest (SW monsoon pollen index) and northeast (NE monsoon pollen index) flow (Table 2, Fig. 2).

Perhaps the most widely used and powerful proxy of past monsoon strength are variations in abundance of the planktonic foraminifer upwelling indicator *Globigerina. bulloides.* It has been shown in the Arabian Sea, and elsewhere at low latitudes, that percentage abundances of *G. bulloides* in surficial sea-sediment are highly and positively correlated with rates of coastal upwelling, nutrient availability, and cooler SSTs (Prell 1984; Prell et al. 1990; Peterson et al. 1997). Because the intensity of upwelling along the Oman continental margin is clearly tied, via Ekman pumping and the positive wind-stress curl, to the strength of monsoon winds, sediment records of *G. bulloides* abundance from this region should be the best monitors of the SW monsoon winds (Prell et al. 1990; Anderson and Prell 1992; Anderson et al. 1992; Anderson 1993).

Observed monsoon variations the past 20000 years

The broad-scale patterns of change recorded in our new high-resolution records of monsoon variability confirm earlier studies of the low-frequency pattern of monsoonal change over the last 20000 year (Fig. 2; Duplessy 1982; Prell 1984; Van Campo et al. 1982; Prell and Van Campo 1986; Sirocko et al. 1991). Abundances of upwelling (monsoon) foraminifera *G. bulloides* are lowest during glacial times, highest during the early-middle Holocene, and then drop off again. This pattern of variation is also supported by our pollen data. Monsoon taxa go from low to high abundances over the last deglaciation, mirrored by decreases in the abundances of pollen indicative of a stronger NE flow from Asia (Fig. 2).

Our new data provide the first available AMS-dated marine pollen and foraminifera records for the fine-scale direct assessment of monsoonal change over the last deglaciation. These records (Fig. 2) confirm the heterogeneous nature of paleo-upwelling records and the corresponding need for more than one sediment core when trying to reconstruct the long-term variation in upwelling, and hence surface wind-stress (Peterson et al. 1991). Our data also confirm that δ^{18}O variations are not a reliable proxy of past monsoon-driven upwelling variations, because they can be influenced, to a varying degree, by changes in global ice volume and variations in regional river discharge. Comparison between our core data, and those of Sirocko et al. (1993), indicate that major shifts in the SW Indian Monsoon can only be inferred reliably when coincident shifts in both *G. bulloides* and the monsoon pollen indices are observed.

Taken together, the three new records suggest four key points regarding monsoon evolution over the last 20000 years. First, the monsoon did not increase monotonically and gradually in response to astronomical forcing over the last deglaciation. Instead, the monsoon appears to have increased in strength abruptly in two steps. The exact timing of these increases are difficult to determine, even with our well-dated records, but they appear to be coincident with the two periods of abrupt δ^{18}O change at about 13 to 12.5 ka (Termination Ia, Duplessy et al. 1981) and between 10 and 9.5 ka (Termination Ib). Second, the monsoon appears to have peaked in strength between 9.5 and 5.5 ka, with the maxima in monsoon pollen probably being the most reliable indicator of peak monsoon conditions because *G. bulloides* percentages appear to saturate above 35–40% (Anderson and Prell 1993). Third, our new high-resolution records confirm that the period of strong monsoon strength (9.5 to 5.5 ka) may have been punctuated by centuries-long periods of weaker monsoon circulation. Finally, our records suggest a close correspondence between the records of monsoon in the Arabian Sea and those from terrestrial sites in the adjacent areas of Africa and Asia. This close correspondence confirms that records from the Oman margin should be useful in reconstructing paleomonsoon variations over much of the late Quaternary (Anderson and Prell 1993).

Although many previously studied marine records of past monsoon variability lack AMS radiocarbon time control, they do have stratigraphic control afforded by δ^{18}O and contain evidence supporting the patterns of change recorded in our new high-resolution records. Both seaward and northward, low-resolution records from the Arabian Sea suggest that peak abundances of *G. bulloides* and monsoon strength occurred in the early Holocene (Prell 1984), whereas to the south and north in the Arabian Sea, marine fossil poll-

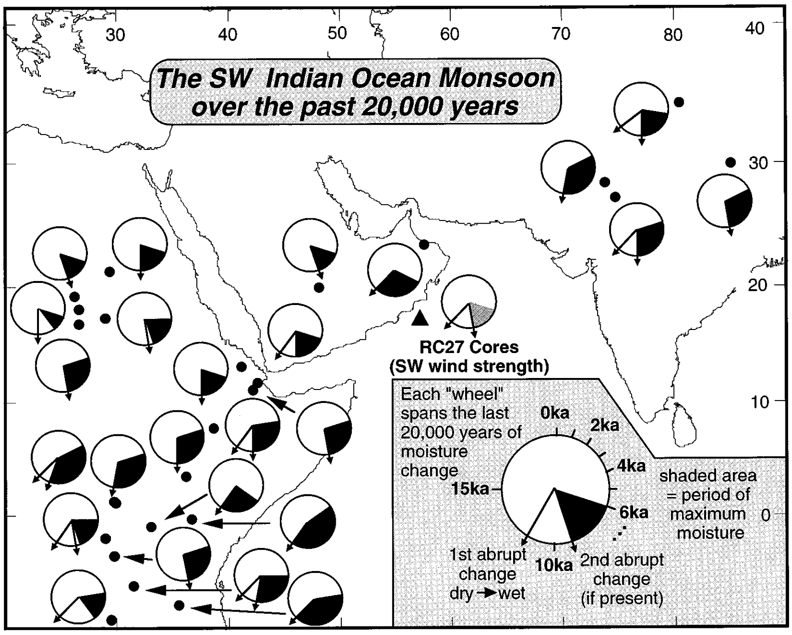

Fig. 3. Map of the Arabian Sea region summarizing changes in the SW Indian Monsoon over the last 20000 years. A "wheel-diagram" is plotted at each terrestrial site (identified in Fig. 1) to summarize the timing of moisture balance changes across the region. All of the sites reviewed were relatively dry compared to today at the last glacial maximum. *Arrows* on each wheel show the timing of abrupt deglacial increases in moisture balance at the site. The *shaded areas* summarize the approximate periods of maximum inferred moisture and monsoon strength at each site. The wheel for our new Arabian Sea records suggests that the timing of abrupt increases in SW monsoon strength, as well as the period of maximum SW monsoon winds, was approximately synchronous with changes over adjacent land areas. All ages are in radiocarbon years BP, with information taken from the literature (Table 1)

en data support a rapid increase in monsoon strength, from weak glacial levels to peak early-Holocene levels, on isotope termination I (Van Campo et al. 1982). Van Campo et al. (1982) also make the observation that centuries-long periods of weakened monsoon strength can be resolved within the hypothesized period of strong monsoonal circulation. None of the previously published marine records of monsoon variation contain a clear picture of change for the period since the early Holocene.

The best-dated previously published marine record of SW Indian monsoon dynamics is one based primarily on a deep water (3212 m) sediment core collected significantly seaward of our cores (Sirocko et al. 1993). Based on an interpretation of oxygen isotopic, dolomite and carbonate variations, Sirocko et al. (1993) suggested that the monsoon increased episodically in a series of four steps between 14.3 and 8.7 ka. Comparison between this data and those presented in this paper, however, indicates that a number of observed abrupt events in each of the marine cores are local in nature and probably not related to changes in the strength of the SW monsoon. Instead, it appears that an abrupt increase in monsoon intensity affected all four of the AMS-dated cores between 13 and 12.5 ka, and then between 10 and 9.5 ka (Sirocko et al. 1993). The entire suite of cores also supports the interpretation that the monsoon was strongest between 9.5 and 5.5 ka (Sirocko et al. 1993).

Terrestrial records of paleomonsoon variability from the African and Asian regions around the Arabian Sea, although not always continuous and well-dated, are sufficient for comparison to the marine record of change in both radiocarbon ka (Figs 1 and 3; Table 2) and in corrected calendar ka (Bard et al. 1990) (Fig. 4, corrected for direct comparison with hypothesized insolation forcing). Consistent with the marine record, the terrestrial records support the concept of a two step increase in monsoon strength during the last deglaciation, and peak monsoon strength in the early to middle Holocene. Almost all of the region surrounding the Arabian Sea was apparently driest just before approximately 13 to 12.5 ka (Fig. 3). Not all records resolve a two-step increase, but those that do suggest that the first increase in monsoon strength occurred at approximately 13 to 12.5 ka (Fig. 4), and the second between 10 and 9 ka. Peak wetness at these sites, scattered from Tibet down through Arabia and East Africa, generally occurred between 10 and 5 ka (about 11.5 and 5.0 calendar ka, Bard et al. 1990), with the end apparently more gradual than the beginning (Fig. 4). Many of the terrestrial records also show evidence of centuries long drier events within the period 13 to 4 ka (Bryson 1989; Gillespie et al. 1983; Street-Perrot and Roberts 1983; Street-Perrott and Perrott 1990; Van Campo and Gasse 1993; An et al. 1993; Roberts et al. 1993; Gasse and Van Campo 1994), quite plausibly corresponding to those suggested by existing and new marine paleomonsoon records (Fig. 2; Van Campo et al. 1982).

Climate forcing and abrupt climatic change

Our new time series of monsoon variation over the past 20 ky, when coupled with existing marine and terrestrial data, suggest a clear pattern of change. Climatic change over the African and Asia regions now influenced by the SW Indian Ocean monsoon appear to have been dominated by this monsoon system over the past 20 ky and probably the entire late Quaternary (Anderson and Prell 1993; Prell and Kutzbach 1992). We confirm that the SW Indian Ocean monsoon was weak until approximately 13 to 12.5 ka, at this time the monsoon strengthened abruptly, increasing rates of coastal upwelling, pollen transport from the SW, and

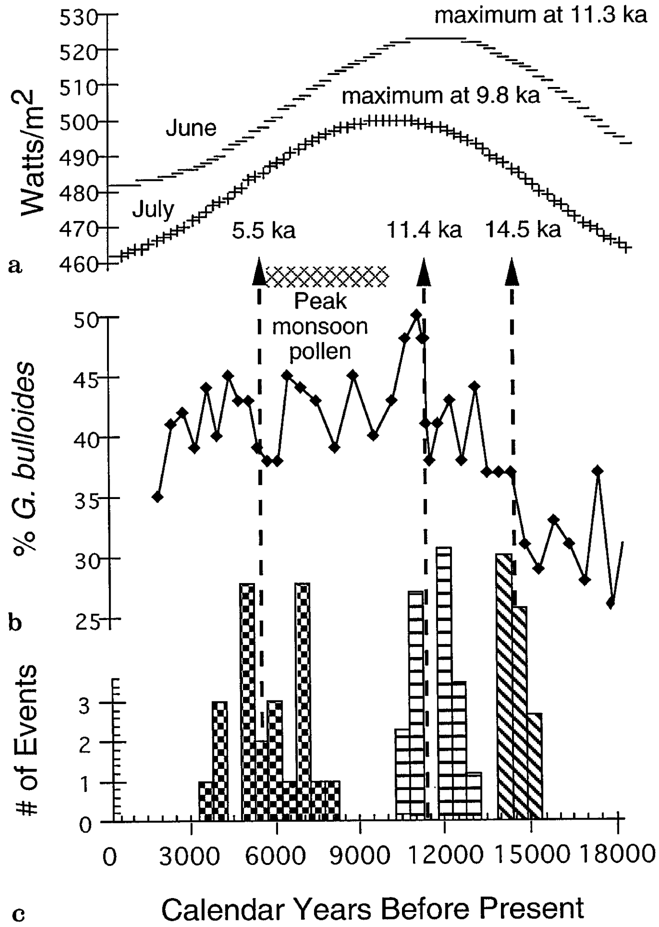

Fig. 4a–c. Summary figure illustrating the hypothesized insolation forcing and the lagged response of the southwest Indian monsoon. In order to compare time series of insolation forcing and monsoon response, all radiocarbon ages in this figure were converted to calendar using the U/Th age calibration of Bard et al. (1992). **a** June and July insolation for 40°N (Berger and Loutre 1991), **b** RC27-23 *G. bulloides* abundance, and **c** histograms of the events mapped in Fig. 3, including the timing of the first abrupt increase in monsoon strength (mean, 14.5 calendar ka), the timing of the second abrupt change (mean, 11.4 calendar ka), and the date at which the period of maximum monsoon strength apparently ended at each site (mean, 5.5 calendar ka). Note that the period of maximum monsoon strength (11 to 5 calendar ka) lags peak insolation (14 to 8 calendar ka). See Figs. 2 and 3 for the timing of events in radiocarbon ka

the advection of moisture over east Africa, Arabia and the Indo-Himalyan region (Sirocko et al. 1993). Our data suggest that a second, stronger increase in monsoon strength occurred between 10 and 9.5 ka, initiating the early to middle Holocene period of maximum monsoon strength in the entire region. Peak monsoon circulation thus lagged peak June insolation by about 3000 calendar years (Fig. 3), before waning throughout the region after 6 ka.

The pattern of Indian Ocean monsoon change we have described seems to correlate surprisingly well with the radiocarbon-dated sequence of abrupt climatic that has been mapped out in the North Atlantic region (Ruddiman 1987; Rind et al. 1986; Atkinson et al 1987; Overpeck et al. 1989; Overpeck 1991; Street-Perrott and Perrott 1991; Birks 1991; Lehman and Keigwin 1992; Lehman and Forman 1992; Svendsen and Mangerud 1992; Karpuz and Jansen 1992; Koc et al. 1993). This leads us to hypothesize that the abrupt changes in monsoon strength, as well as the 3 ka lag in response to astronomical forcing, is due primarily to the influence of abruptly changing glacial climate boundary conditions over the last 15 ky. If insolation alone was driving the monsoon, monsoon strengths would have increased gradually in parallel with summer insolation over the latest Pleistocene to a maximum at 10 ka. Instead, monsoon strength increased abruptly between 13 and 12.5 ka, at the same time that the North Atlantic polar front shifted abruptly northward, warming the northeast Atlantic region as far north as at least 80°N (Ruddiman 1987; Bard et al. 1987; Lehman and Forman 1992; Svendsen and Mangerud 1992). The even larger abrupt increase in monsoon strength between 10.0 and 9.5 ka coincides with the largest and most rapid warming of the North Atlantic, as well as the rapid final melting of European ice sheets (Rind et al. 1986; Ruddiman 1987; Street-Perrott and Perrott 1991; Lehman and Keigwin 1992; Lehman and Forman 1992; Svendsen and Mangerud 1992). North Atlantic temperatures as far north as 80°N reached their interglacial highs, and were warmer than present, between 9.5 and 4.0 ka (Birks 1991; Svendsen and Mangerud 1992).

Maximum monsoon strength and North Atlantic warmth were thus coincident with each other between 9.5 and 5.5 ka, lagging significantly the period of maximum summer insolation. Furthermore, peak monsoon strength and North Atlantic warmth began abruptly at 9.5 ka, and waned more gradually after 5.5 ka. We hypothesize that monsoon strength increased non-linearly to insolational forcing until the influence of glacial boundary conditions, primarily cold North Atlantic SSTs and European ice sheets, abruptly declined between 10.0 and 9.5 ka. After 5.5 ka, the monsoon weakened more gradually in concert with slowly decreasing summer insolation.

Physical climatology and climate model experiments provide an idea of how the observed changes in monsoon strength may have been linked to abrupt changes in the North Atlantic and European regions. Abrupt warming in the North Atlantic at 13 to 12.5 ka may have led to slightly greater warming over the Tibetan Plateau in spring and summer, a response supported by climate model experiments (Rind et al. 1986). This warming may have been sufficient to reduce Tibetan Plateau snow cover in spring and early summer, and hence (via influences on albedo and soil hydrology) to permit the development of a warmer Tibetan Plateau and an earlier and stronger monsoon (Duplessy 1982; Dickson 1984; Barnett et al. 1989; Sirocko et al. 1991). An abrupt increase in atmospheric CO_2 and CH_4 concentrations at the same time (Barnola et al. 1987; Chappellaz et al. 1990; White et al. 1994) may have also acted in summer to warm the Tibetan Plateau more than the ocean, and thus drive a stronger mon-

Table 3. Climate model experiment designation and description

Experiment name	Description
WARM	• 11 ka orbital parameters (Berger and Loutre 1991) • 11 ka land ice (Denton and Hughes 1981), with the exception that the Scandinavian Ice Sheet was removed • Current sea surface temperatures (SSTs)
COLD	• 11 ka orbital parameters (Berger and Loutre 1991) • 11 ka land ice (Denton and Hughes 1981), including the Scandinavian Ice Sheet • Current SSTs, except in the North Atlantic, where 18 ka SSTs were prescribed north of 25°N (CLIMAP 1981)

soon (Prell and Kutzbach 1992; Meehl and Washington 1993). This change may have, in turn, triggered land cover (i.e., vegetation) changes that also contributed, via positive feedbacks, to a stronger monsoon (Street-Perrott et al. 1990; Overpeck 1993; Gasse and Van Campo 1994).

We recently performed two additional climate model sensitivity experiments to investigate how abrupt change in the North Atlantic-European region might have affected the strength of the SW Indian Monsoon. Following the experimental strategy of Rind et al. (1986; see also Overpeck et al. 1989), the two experiments, carried out with the Goddard Institute for Space Studies (GISS) Model II general circulation model (Hansen et al. 1983). The model simulates the full seasonal cycle, solving the equations for conservation of mass, energy, momentum, and moisture, and calculates the radiative fluxes, cloud cover, surface fluxes, etc. (Rind et al. 1986). It has realistic topography at an $8° \times 10°$ (latitude by longitude) horizontal resolution with fractional grid representation for land and ocean. Ground temperature calculations include the diurnal variation and seasonal heat storage, while ground hydrological parameters are a function of vegetation type. Hansen et al. (1983) has shown that the model produces realistic temperature fields when modern sea surface temperatures (SSTs) are prescribed. Our two new experiments were identical except for two important differences (Table 3; Fig. 5). Both experiments were run for five years with the same 11 ka insolation, as well as prescribed SSTs and ice-sheet configurations that were identical except in the North Atlantic-European region. In this region, one experiment (WARM) was prescribed to have modern SST's and no Scandinavian Ice Sheet, whereas the other experiment (COLD) was set to have glacial (Younger Dryas) SSTs down to about 40°N, as well as a Scandinavian ice sheet with approximately 11 ka height and extent (Fig. 5; Rind et al. 1986). We used 11 ka boundary conditions to be comparable to previous model results (where only SSTs were changes), and because it

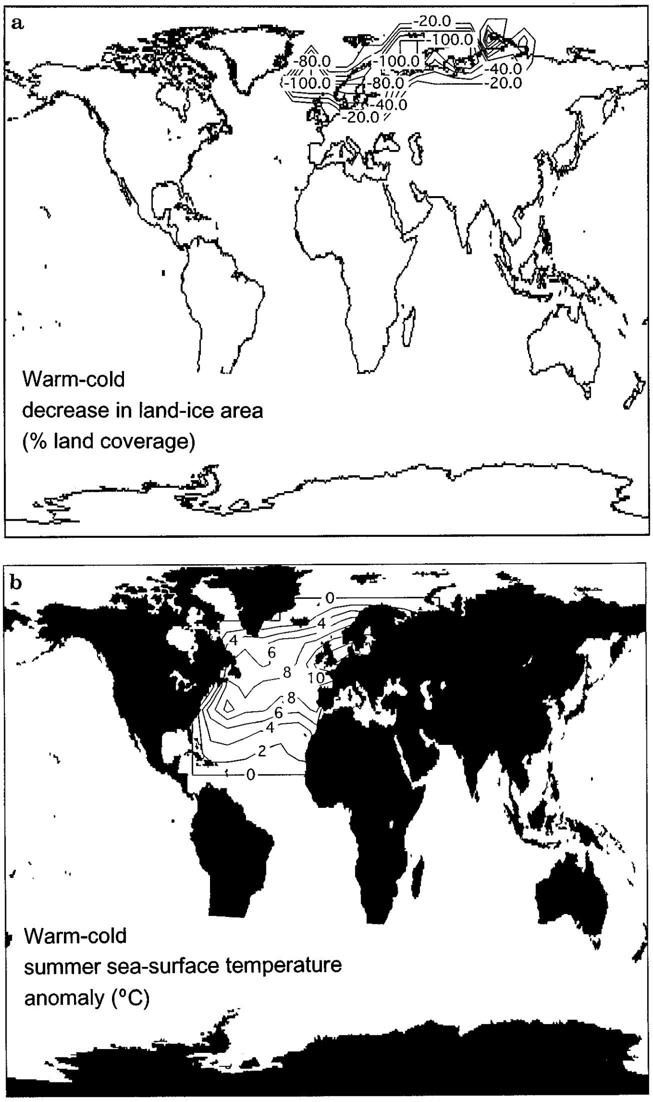

Fig. 5. a Maps illustrating prescribed boundary condition sea-surface temperature (SST), and **b** ice-sheet differences, between the two climate model simulations (WARM experiment-COLD experiment; Table 3). The WARM experiment SSTs for each month were prescribed to be the same as modern for all ocean areas, whereas the COLD experiment was the same, except in the North Atlantic, where SSTs were prescribed to be equivalent to their glacial (Younger Dryas) values down to approximately 25°N (Rind et al. 1986). In addition, experiment WARM did not have any Scandinavian Ice Sheet, whereas experiment COLD had a Scandinavian Ice Sheet with approximately 11 ka height and extent (Table 3; Rind et al. 1986). Both experiments were run for five years, and the averages for the last four years were compared (e.g., Fig. 6)

would not decrease the applicability of the climate model sensitivity experiments to the 10 to 9.5 ka event (Rind et al. 1986; Overpeck et al. 1989).

The two climate model simulations were run to approximate the abrupt change that apparently took place in the North Atlantic-European region between 10 and 9.5 ka. Given the observation that early Holocene high-latitude North Atlantic temperatures were warmer than present, and that the Scandinavian Ice Sheet disappeared significantly before the Laurentide

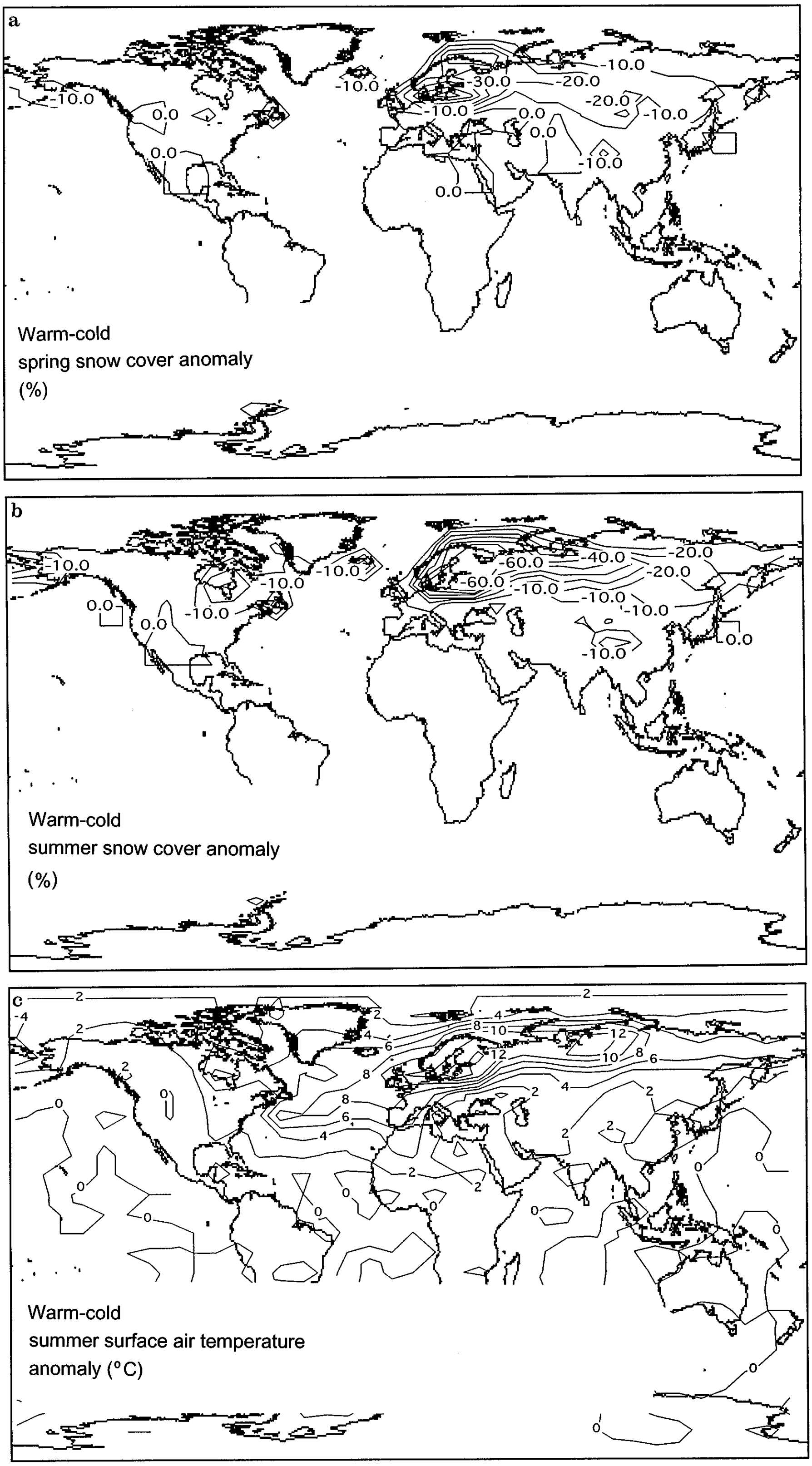

Fig. 6a–c. Simulated spring and summer anomalies between climate experiment WARM minus experiment COLD. These anomalies illustrate the type of changes that would have been induced by an abrupt warming of the North Atlantic region, coupled with a rapid disappearance of the Scandinavian Ice Sheet. These abrupt changes resulted in a significant warming and **a** a decrease in snow cover on the Tibetan Plateau during spring, as well as significant summer **b** snow cover reduction and **c** warming. These changes would have led to an earlier and stronger SW Indian monsoon circulation

Ice Sheet (Birks 1991; Svendsen and Mangerud 1992), it is likely that our prescribed North Atlantic SST increases (i.e., the difference between SSTs in the COLD and WARM experiments) were an underestimate of the warming that actually took place between 10 and 9.5 ka. Comparison between the two climate model simulations (WARM-COLD) suggests the type of climatic change that may have taken place over the Tibetan Plateau in response to the abrupt changes in SST and ice-sheet height/extent upstream and to the west.

Whereas earlier 11 ka model results (Rind et al. 1986) suggest that the SST changes alone were not enough to drive a shift in monsoon strength, our new results (Fig. 6) indicate that these abrupt North Atlantic-European changes together may have been sufficient to significantly warm the Tibetan Plateau in all seasons (only spring and summer are plotted), and to cause an earlier melting of snow cover in spring. This reduction in spring snow cover, coupled with greater warming of the Plateau in summer (Fig. 6) would have led to an abruptly stronger monsoon circulation, with a resulting increase in available moisture over much of the area affected by this monsoon.

Our summary of the available data (Figs. 3 and 4) suggests that, whereas the deglacial intensification of the monsoon was probably concentrated in two abrupt events, the weakening of the monsoon during the middle to late-Holocene was apparently more gradual, and less concentrated into any single event. This assertion makes sense in light of the fact that the influence of glacial boundary conditions on the Tibetan Plateau disappeared after the early Holocene, allowing the strength of Plateau heating to be more directly linked to Milankovich-driven insolation forcing. As summertime Northern Hemisphere insolation waned after the early Holocene, the monsoon gradually weakened in strength.

Despite the assertion made above, many sites around the monsoon region (Fig. 1) apparently did experience a sharp reduction in effective moisture between 7 and 3 ka. It is possible that these changes were influenced by forcing other than Milankovitch alone. In addition, many sites throughout the region, including those represented by our new Arabian Sea data, support the existence of significant century-scale monsoon weakening events during the early to middle Holocene (Van Campo et al. 1982; Bryson 1989; Gillespie et al. 1983; Street-Perrott and Roberts 1983; Street-Perrott and Perrott 1990; Van Campo and Gasse 1993; An et al. 1993; Gasse and Van Campo 1994). Separate climate model results suggest that possible thermohaline-driven changes in North Atlantic sea-surface temperatures during the Holocene were probably not sufficient to affect the Tibetan Plateau (Rind and Overpeck 1993). In contrast, if recent assertions that solar irradiances may have varied by as much as 25% during the Holocene are correct (Lean et al. 1992; Lean and Rind 1994), then it is quite plausible that the Tibetan Plateau may have cooled by as much as 1 °C for decades or centuries (Rind and Overpeck 1993). Decades to centuries long changes in tropospheric aerosol loading (e.g., Zielinski et al. 1994) or trace-gas concentrations (e.g., White et al. 1994), may have also been the cause of possible decade to century-scale abrupt changes in monsoon intensity (Bryson 1989). Our summary of available paleomonsoon time series, however, indicates that these series are not well enough dated to define the exact patterns of decade to century-scale variations in past monsoon strength, nor to identify their causes.

Conclusions

The comparison of our new Arabian Sea time series with terrestrial records of monsoon change over Africa and Asia supports the idea that the southwest Indian monsoon did not respond linearly to insolation forcing over much of the past 20 ka. Instead, reductions of glacial climate boundary conditions at 13 to 12.5 ka and 10 to 9.5 ka (about 15.3 to 14.7 and 11.5 to 10.8 calendar ka, Bard et al. 1990) may have acted as triggers in permitting the Tibetan Plateau to warm strongly in the summer and generating the pressure gradient needed for a strong southwest Indian monsoon. After the early Holocene and disappearance of glacial boundary conditions, millennial-scale variations in the SW monsoon appear to have been linked more linearly to insolational forcing, particularly as the monsoon weakened gradually after 5 ka. Although abrupt changes in North Atlantic SST, northern European ice cover, and atmospheric trace-gases are coincident with the two abrupt increases in monsoon strength, it is also plausible that abrupt changes in Southern Hemisphere trade winds may have affected deglacial monsoon strength (Clemens and Prell 1990, 1991; Clemens et al. 1991; Clemens and Oglesby 1992). Whereas the trade wind hypothesis may be difficult to test due to the lack of sufficiently high-resolution southern Indian Ocean sediment cores, it should be possible to determine if and when the Tibetan Plateau warmed abruptly, using terrestrial evidence from the Tibetan Plateau. Available data suggest that the Tibetan Plateau may have warmed abruptly at the same times that the monsoon increased in strength, particularly between 10 and 9.5 ka, and that these abrupt changes were coincident with rapid decreases in snow and glacier coverage (Wang and Fan 1987; Fang 1991; Gasse et al. 1991; Jarvis 1993; Porter et al. 1992; An et al. 1993; Gasse and Van Campo 1994). However, more well-dated paleoclimatic records from the Tibetan Plateau are needed to fully test our hypothesis.

Large abrupt changes appear to have taken place throughout the climate system over the last deglaciation, and the southwest Indian monsoon is no exception. Our results illustrate a mechanism by which abrupt change in the North Atlantic region may have driven abrupt change a quarter of the way around the globe, and at low, even southern latitudes. Climate model experiments seem to confirm that deglacial climate change over NW Africa may have also been linked to the same high-latitude abrupt change (Fig. 6; Overpeck et al. 1989; Street-Perrott and Perrott 1990; Lamb et al. 1995). Our new Arabian Sea data support the previous finding that major century-scale changes in monsoon strength were common during the present interglacial (Bryson 1989; Gillespie et al. 1983; Street-Perrott and Roberts 1983; Street-Perrott and Perrot 1990; Van Campo and Gasse 1993; An et al. 1993; Gasse and Van Campo 1994). Multiple well-dated records of past monsoon change are needed to define the exact nature and timing of these century-scale events, an exercise that may be crucial to understand how the monsoon may change in the future (Broecker 1987). This is particularly true since the past abrupt events occurred when Northern Hemisphere summers were significantly warmer than present (COHMAP members 1988), just as future summers may be warmer than present in a greenhouse-warmed world (Houghton et al. 1992).

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