JOURNAL OF QUATERNARY SCIENCE (2010) **25**(6) 911–917
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Published online 5 February 2010 in Wiley InterScience
(www.interscience.wiley.com) DOI: 10.1002/jqs.1369

Indian summer monsoon during the last two millennia

DAVID M. ANDERSON,[1*,†] CORINNE K. BAULCOMB,[1] ALICE K. DUVIVIER[1] and ANIL K. GUPTA[2]
[1] Paleoclimatology Branch of NOAA's National Climatic Data Center, Boulder, Colorado, USA
[2] Department of Geology and Geophysics, Indian Institute of Technology, Kharagpur, India

Anderson, D. M., Baulcomb, C. K., Duvivier, A. K. and Gupta A. K. 2010. Indian summer monsoon during the last two millennia. *J. Quaternary Sci.*, **Vol. 25** pp. 911–917. ISSN 0267-8179.

Received 16 April 2009; Revised 27 October 2009; Accepted 12 November 2009

ABSTRACT: The monsoon is a large-scale feature of the tropical atmospheric circulation, affecting people and economies in the world's most densely populated regions. Future trends due to natural variability and human-induced climate changes are uncertain. Palaeoclimate records can improve our understanding of monsoon dynamics and thereby reduce this uncertainty. Palaeoclimate records have revealed a dramatic decrease in the Asian summer monsoon since the early Holocene maximum 9 ka BP. Here we focus on the last 2 ka, where some records indicate an increasing trend in the summer monsoon. Analysing *Globigerina bulloides* upwelling records from the Arabian Sea, we find the weakest monsoon occurred 1500 a BP, with an increasing trend towards the present. Copyright © 2010 John Wiley & Sons, Ltd.

KEYWORDS: monsoon; Holocene; Tropics; India; Arabian Sea.

Introduction

Earth's climate is changing as a result of both natural and anthropogenic causes. One region of the world that is particularly susceptible to shifts in climate and extreme weather events is the Indian subcontinent. In this densely populated region both droughts and floods, as well as trends in the seasonal receipt of rainfall, have the potential to be devastating from economic, agricultural and public health perspectives. The Southwest Indian Monsoon is the main contributor to India's yearly rainfall budget. Therefore, unravelling past trends of the monsoon rains is critical to understanding future monsoon trends.

India is home to one of the longest time series of instrumental climate observations, the all-India monsoon rainfall index, spanning 1844 to the present (Sontaake *et al.*, 1993; Parthasarathy *et al.*, 1994). This record is remarkable both for showing an absence of a long-term trend and for showing a large variation in rainfall from one year to the next. Records of the atmospheric circulation such as the Indian Monsoon Index (IMI), based on the Arabian Sea winds, are much shorter and span only the past 50 a (Wang *et al.*, 2001). These records also indicate large variations from one year to the next and do not reveal a trend. Blending appropriate instrumental records with palaeo observations, from ocean sediments, cave deposits or other proxies, can provide a longer perspective on the monsoon and its natural variability than otherwise would be available.

The monsoon is the result of both the seasonally changing thermal contrast between the land and ocean, and the seasonal migration of the Intertropical Convergence zone (ITCZ), which reaches its northernmost location in summer, exacerbating the thermally driven pressure gradient (Gadgil, 2003). During winter, the air over the land is cooler and denser than the air over the water. This creates a surface pressure contrast that is enhanced by the position of the ITCZ over the ocean south of the continent of India, resulting in a dry northeasterly wind that blows across the continent. In spring the continent warms quickly compared to the surrounding water. The winds then reverse, and a southwesterly wind blows across the Arabian Sea, bringing moisture-laden air towards India. The southwesterly winds that bring the monsoon rains also cause upwelling and sea surface cooling in the Arabian Sea, producing a distinctive fossil record that can be tied to the monsoon winds via well-known physical mechanisms.

A variety of Indian and Asian summer monsoon indices have been developed from instrumental records. In general, throughout India and Asia in summer, stronger winds over the ocean and increased rainfall over land are correlated with the large-scale atmospheric circulation (Fig. 1). The Webster–Yang Index (WYI) is a broad-scale index of the monsoon circulation, covers a large continuous area (40–110° E, 0–20° N), and measures the difference in strength between low-level (850 hPa) and upper-level (200 hPa) winds (Webster and Yang, 1992). In contrast, the IMI measures the Arabian Sea winds by subtracting the low-level (850 hPa) zonal wind over land (20–30° N, 60–90° E) from the low-level (850 hPa) winds in

* Correspondence to: D. M. Anderson, Paleoclimatology Branch of NOAA's National Climatic Data Center, 325 Broadway, E/CC23, Boulder, CO 80305, USA.
E-mail: david.m.anderson@noaa.gov
† The contribution of D. M. Anderson to this article was prepared as part of his official duties as a United States Federal Government employee.

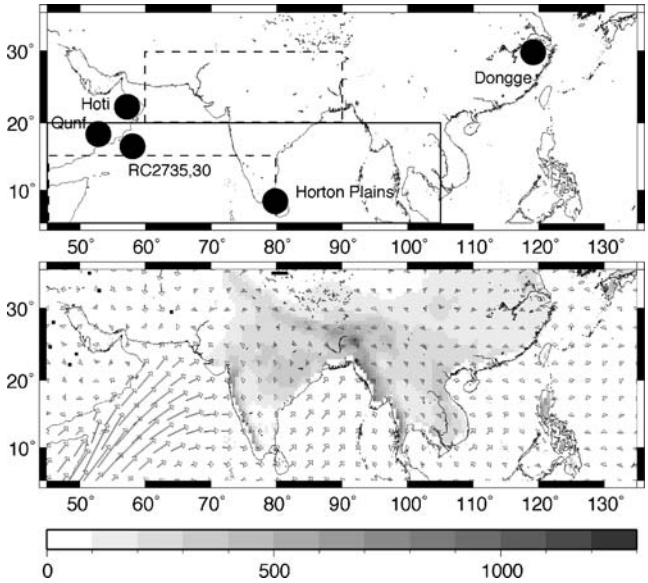

Figure 1 Modern July wind stress and precipitation, location of other records containing a weaker-than-present interval during the late Holocene (top panel), and the Index regions described in Table 1 (top panel)

the Arabian Sea (5–15° N, 40–80°E) (Wang *et al.*, 2001). The year-to-year correlation among these indices for June, July and August for the period 1982–2004 shows that the strongest correlation is between the Arabian Sea winds (IMI) and all-India monsoon rainfall (AIMR) (0.65) (Table 1). The second strongest (negative) correlation is between the Arabian Sea winds (IMI) and Arabian Sea sea surface temperature (SST), expected given the physics of upwelling in the Arabian Sea. The summer SST is negatively correlated with all three monsoon circulation indices over the 22-year period. A more extensive comparison of Indian summer monsoon (ISM) indices supports the correlations between winds and rainfall identified here, and describes their independence from a centre of convection formed by the South East Asian summer monsoon (SEASM) (this region includes the Philippine Sea) (Wang and Fan, 1999). Generally, the correlations between the ISM indices and the SEASM indices never exceed 0.33. For some indices the correlations are weakly negative (−0.06 to −0.23). Based on these correlations from instrumental data, one might expect palaeo proxies from the Arabian Sea, sensitive to the ISM, to correlate only weakly to palaeo proxies from cave deposits in China that are sensitive to the SEASM.

Table 1 Correlation among monsoon indices and sea surface temperature in the northwestern Arabian Sea during June–July–August for the period 1982–2004

Index	WYI	IMI	AIMR	SST
WYI	1.0			
IMI	0.21	1.0		
AIMR	0.21	0.65	1.0	
SST	−0.12	−0.31	−0.24	1.0

WYI, Webster–Yang Index, defined as the zonal wind shear U850–U200 (0–20°N, 40–110° E), where U850 denotes the zonal wind at 850 hPa height; IMI, Indian Monsoon Index, defined as U850 (5–15°N, 40–80° E)–U850 (20–30° N, 70–90° E); AIMR, All-India Rainfall Index (Parthasarathy *et al.*, 1994); SST, June–August sea surface temperature averaged for 16–19° N, 56–59° E determined from the Optimally Interpolated Sea Surface Temperature Data Set Version 2 (Reynolds *et al.*, 2002).

Palaeo proxies of the monsoon have been developed from Arabian Sea sediments and more recently from cave deposits and other terrestrial records (Prell, 1984; Neff *et al.*, 2001; Fleitmann *et al.*, 2003; Gupta *et al.*, 2003a). All records are consistent in indicating a stronger Indian summer monsoon during the early Holocene 9 ka ago. This maximum has been attributed to the alignment of perihelion with the summer solstice that increased summer radiation by 8% relative to the present (Kutzbach, 1981). Detection of the Holocene monsoon maximum across Asia in many different regions affected by the monsoon (Morrill *et al.*, 2003), and the increase in winds and rainfall in numerical models in response to the prescribed increased radiation, is one of the significant advances in Quaternary palaeoclimatology (Kutzbach and Otto-Bliesner, 1982; Kutzbach and Guetter, 1984). While various proxies have different advantages, proxies based specifically on upwelling in the Arabian Sea remain attractive for quantitative reconstructions because of the direct, physical link between upwelling and the surface wind. Furthermore, the abundance of the planktic foraminifer, *Globigerina bulloides*, is exclusively linked to the upwelling process, making *G. bulloides* an ideal subject for palaeo proxies. *G. bulloides* is considered a subpolar species and, without the upwelling and cooling caused by the monsoon, would be absent from the tropical waters of the Arabian Sea (Prell, 1984). In numerous studies (Prell, 1984; Overpeck *et al.*, 1986; Naidu and Malmgren, 1996; Anderson *et al.*, 2002; Gupta *et al.*, 2003), the percentage of *G. bulloides* was measured relative to all other species present as a means to reconstruct past trends in monsoon intensity. In Anderson *et al.* (2002), the trend in southwestern monsoon intensity was investigated with fine-resolution (2 mm interval) sediment core analysis. The record revealed that a minimum in monsoon intensity occurred approximately 400 a BP and that the past 400 a have witnessed a dramatic increase in monsoon intensity (Anderson *et al.*, 2002). The purpose of this study is to improve the dating and extend the high-resolution analysis performed on sediment cores taken from the Oman Margin by Anderson *et al.* (2002) with the goal of understanding changes in monsoon intensity that occurred over the past 2 ka.

Methods

Soutar box subcores taken from sites RC2730 (18° 13′ N, 57° 41′ E) and RC2735 (18° 14′ N, 57° 36′ E) were sampled by dividing one half of the subcore into continuous 2 mm intervals from a depth of 100–200 mm. The individual sediment samples were freeze-dried, weighed, soaked in tap water overnight and washed through a 150 μm sieve. The sieved foraminifer shells were dried, weighed for percentage calcium carbonate coarse fraction and stored for counting. Additionally, the first litre of sediment wash from each sample was collected and dried in order to collect clay, pollen and juvenile foraminifers.

The coarse fraction samples were first split in half to create an archive half for counting, and the other half for radiocarbon and other analyses. The archive was split repeatedly to produce an aliquot containing at least 300 shells that were placed on a paper micropalaeontology slide. The samples comprising each core were randomly counted by one of three individuals and following the initial counts some samples were recounted to evaluate accuracy and precision. The interval 0–100 mm of RC2735 was counted by D. Anderson and 0–100 mm of

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RC2730 was counted by A. Gupta as part of the earlier study (Anderson *et al.*, 2002).

Radiocarbon was measured either on *G. bulloides*, on *G. bulloides and N. dutertrei*, or mixed planktonic foraminifers (*G. bulloides*, *G. glutinata*, *N. dutertrei* and *G. ruber* dominate the assemblage) where insufficient *G. bulloides* were found, combining samples from adjacent depth levels when necessary to obtain sufficient material. Species and depth intervals analysed appear in Table 2. All samples were hand-picked and sonified in distilled water. The measurements in Anderson *et al.* (2002) were made at the National Ocean Sciences Accelerator Mass Spectrometry (NOSAMS) Facility (OS numbers), and measurements for this study were made at the University of Arizona (AA numbers) (Table 2). The reliability of ages determined on foraminifers from upwelling regions and from deeper-dwelling foraminifers can be questioned. Deep water is older with lower radiocarbon activity, and this bias might appear in upwelling regions where deep water upwells or in deep-dwelling foraminifers. Two observations indicate this problem is unlikely. First, samples from the top of the box cores have the same age as that expected for the sea surface (435 a) or even younger ages, which we attribute to the presence of bomb radiocarbon. Second, a *t*-test for the difference in ages between the surface dweller (*G. bulloides*) and the deep dweller (*N. dutertrei*) fails to reject the null hypothesis (*N. dutertrei* age not older than *G. bulloides* age) based on the ages of seven paired samples ($t = 0.67$ assuming unequal variances between the samples, d.f. $= 6$, *t*-critical $= 1.95$ for chance of type 1 error < 0.05, one-tailed probability $= 0.26$) (Anderson *et al.*, 2002). One explanation is that the upwelled water comes from the upper few hundred metres where the radiocarbon activity is not much lower than the surface ocean.

Results

New radiocarbon measurements made on both box subcores improve the stratigraphy published in Anderson *et al.* (2002) and extend both records farther back in time (Table 2 and Fig. 2). The sediment in both cores is homogeneous nannofossil-rich foraminifer ooze, without lamination, and without mottling or other evidence of bioturbation, consistent with deposition in the oxygen minimum zone of the northwestern Arabian Sea. The radiocarbon ages were converted to calendar years before 1950 AD using the Calib rev. 5.0.2 program, and assuming a constant surface ocean reservoir correction (*R*) of 207 (Stuiver and Reimer, 1993). The calibrated age for both cores is less than zero at the top, indicating the presence of bomb radiocarbon. Consequently, we assigned the date of collection (1986) to the top of the cores. Linear age–depth relationships for both cores, as shown in Fig. 2, were constructed, from which the sedimentation rate was extrapolated. Small age reversals were found towards the bottom of both cores. One other age reversal exists in RC2730 between adjacent samples measured in the original and subsequent study. Here, we also used the extrapolated sedimentation rate. While the radiocarbon uncertainty is around 50 a for these samples (Table 2), the uncertainty due to changes in the surface ocean reservoir age probably creates an age uncertainty of 100 a or more for the past 6 ka.

An unexpected result was that the sedimentation rate is five times slower in RC2735 (3.3 cm ka^{-1}) compared to RC2730 (20 cm ka^{-1}). This changes the stratigraphic correlation made by Anderson *et al.* (2002) and indicates that the 2 mm sample resolution in RC2735 is 60 a, compared to the 2 mm resolution of 10 a in RC2730. Combining data from these two cores is therefore unreasonable because of the different resolution. Other implications of the revised correlation are discussed below.

We counted the percentage of *G. bulloides* between 100 and 200 mm in both cores, extending the results from the previous study (Anderson *et al.*, 2002) and encountering several surprises. Connected to the unanticipated difference in sedimentation rates is the issue of stratigraphic correlation. The two *G. bulloides* stratigraphies appeared to correlate well in the depth domain, beginning with high percentages at the core top, decreasing to minima (40 mm level in RC2730, 60 mm level in RC2735), and returning to similarly high percentages at 200 mm. However, as stated previously, the

Table 2 Radiocarbon measurements for cores RC2735 and RC2730 from this study and Anderson *et al.* (2002). Calibration to years before AD 1950 AD used Calib rev. 5.0.2 and $R = 207$ (Stuiver and Reimer, 1993)

Core	Depth (mm)	Depth range (mm)	Species	Radiocarbon age (a BP)	Calibrated radiocarbon age (a BP)	1σ (a)	Lab. #
RC2730	1.5	0–2	Mixed	325	—	40	OS-33148
RC2730	1.5	0–2	Mixed	250	—	30	OS-33170
RC2730	5	4–6	*G. bulloides*	345	—	30	OS-33149
RC2730	9	8–10	*G. bulloides* and *N. dutertrei*	450	—	30	OS-33150
RC2730	30	28–32	*G. bulloides* and *N. dutertrei*	885	321	30	OS-33151
RC2730	39	38–42	*G. bulloides* and *N. dutertrei*	1000	446	30	OS-33152
RC2730	48	46–50	*G. bulloides* and *N. dutertrei*	970	412	30	OS-33167
RC2730	73	70–76	*G. bulloides* and *N. dutertrei*	1280	574	55	OS-33168
RC2730	98	96–100	*G. bulloides* and *N. dutertrei*	1520	861	30	OS-33169
RC2730	100	100–101	Mixed	1137	526	34	AA67288
RC2730	138	138–139	*G. bulloides* and *N. dutertrei*	1541	876	36	AA67289
RC2730	174	174–175	Mixed	1923	1270	35	AA67290
RC2730	196	196–197	Mixed	1680	1010	36	AA67291
RC2735	0	0–10	*G. bulloides*	435	—	55	OS-16706
RC2735	102	102–103	Mixed	2844	2330	42	AA67292
RC2735	128	128–129	*G. bulloides* and *N. dutertrei*	3747	3418	40	AA67293
RC2735	166	166–167	*N. dutertrei*	5211	5359	41	AA67294
RC2735	198	198–199	*N. dutertrei*	5244	5380	59	AA67295

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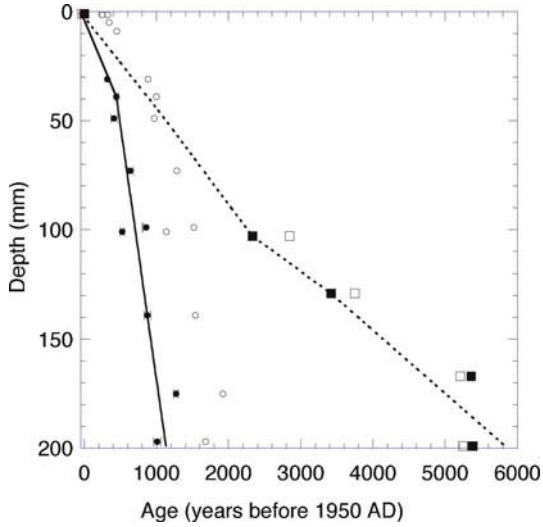

Figure 2 Age–depth relationship for box subcores RC2730 (circles) and RC2735 (squares), including the original (open symbol) and calibrated (filled symbol) radiocarbon ages including 1σ error (vertical lines)

radiocarbon dates do not support this correlation. Radiocarbon dates indicate that the level of 200 mm in RC2730 correlates with 50 mm in RC2735. The second surprise is the abrupt change in abundance found in RC2730 at 100 mm. The new percentages (below 100 mm in depth) counted are 20% higher than the counts made by A. Gupta in the original study (0–100 mm). Possible explanations are that (1) differences in the sample washing altered the preservation of *G. bulloides*, or we (D. Anderson, C. D'Ippolito, A. DuVivier) were more inclusive in our taxonomic recognition of *G. bulloides* compared to A. Gupta. Another core (Hole 723A) analysed by A. Gupta also has lower percentages compared to the Holocene minimum and maximum; however the Hole 723A counts can be scaled to agree with the Holocene minimum and maximum observed in other cores (Prell, 1984; Anderson and Prell, 1993) by multiplying the percentage data by 1.33 and adding 12%. We scaled the RC2730 0–100 mm counts by the same amount.

The resulting time series of *G. bulloides* contains the following events (Fig. 3). Between 6000 and 1500 a BP, there is a decreasing trend in abundance, with a minimum around 1500 a BP. Following this minimum, abundance increases towards the present, with a brief maximum during 1200–800 a BP, a minimum 400 a BP, followed by an increasing trend towards the present. These events are found in our previous study (Anderson *et al.*, 2002) as well as the analysis of Hole 723A (Gupta *et al.*, 2003b). We are reluctant to evaluate smaller-scale events and century-scale variability in these records given the large between-core differences (RC2730, RC2735, and Hole 723A) and the many potential sources of uncertainty stemming from these differences.

Discussion

G. bulloides production and relationship to monsoon wind speed

Sediment trap time series reveal that approximately 73% of the total foraminifer flux in the northwestern Arabian Sea occurs during the months of July to October (Curry *et al.*, 1992). August plankton tows contain two orders of magnitude higher concentrations of *G. bulloides* compared to February tows (Peeters *et al.*, 2002). Away from the upwelling region, and at different times in the past, this useful seasonal bias in production may change, but for the past few thousand years we hypothesise that the Oman Margin sediments reflect the conditions during July–October.

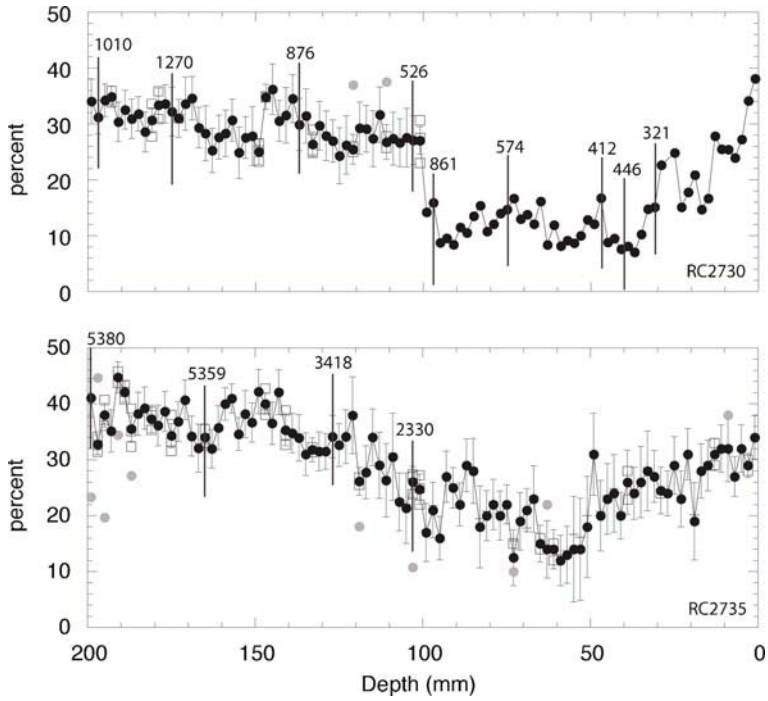

Figure 3 *G. bulloides* abundance (%) relative to all planktic foraminifers in core RC2730 (top) and RC2735 (bottom), Agresti Coull confidence interval (95%) (grey bars), duplicate counts (squares) and rejected counts (grey circles)

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During the summer months, stronger southwestern monsoon winds and increased rainfall correlate with cooler SST, as noted in the Introduction. Several lines of evidence show that the percentage of *G. bulloides* relative to all other species increases with decreasing SST. One observation is the geographic correlation between the abundance of *G. bulloides* in core top sediments and the SST; the greatest percentage of *G. bulloides* coincides with the coolest SST (Prell, 1984). A second observation is that the percentage of *G. bulloides* in the sediment trap samples increases as the SST cools throughout the upwelling season (Curry *et al.*, 1992). We updated the linear relationship found by Prell using additional core top samples in order to reconstruct SST from *G. bulloides* (SST°C $= -0.05$ (*G. bulloides* %) $+ 28.1$°C, $R^2 = 0.42$), and then used the relationship between SST and the IMI monsoon wind index observed during the period 1984–2004 to reconstruct the IMI (IMI (m s^{-1}) $= -0.31 \times \mathrm{SST}_{\mathrm{jja}} + 15.41$, $n = 22$, $R^2 = 0.1$. This provides a preliminary reconstruction of the IMI from fossil *G. bulloides* (Fig. 4, right axis), facilitating comparison with other records; however the low R^2 indicates that more work is needed to refine this reconstruction. We expect other geological records influenced by wind and precipitation to be correlated with each other in the past because the correlations shown in Table 1 show that the Arabian Sea winds (IMI) correlate with the large-scale monsoon circulation (WYI). Both of the wind indices correlate with increased rainfall, and all three atmospheric measures correlate with cooler SST.

Comparison with other records

Monsoon records across Asia reveal that the summer monsoon weakened between 9 ka BP and the present. Some, but not all, indicate that the weakest monsoon occurred prior to today, resulting in an increasing trend over the last 1–2 ka. Records containing this minimum include the Dongge Cave oxygen isotope records from China (Wang *et al.*, 2005) (Fig. 1) and the Qunf and Hoti Cave records from Oman (Fleitmann *et al.*, 2003). The Dongge Cave record reveals minima at 1600 a BP and 500 a BP, similar to the Arabian Sea record. The similarity between Indian (RC2730, RC2735, Qunf, Hoti) and East Asian monsoon (Dongge) records is surprising given the generally low correlation between ISM and SEASM indices described in the Introduction. In the Qunf cave, reinitiation of growth in the recent past beginning around 600 a BP and depletion in oxygen isotopes parallel the increase in *G. bulloides* percentages during the past centuries. The more finely resolved cave records show a weaker monsoon during the Little Ice Age 500 a ago, and a stronger monsoon during the Medieval Warm Period (AD 900–1300) that is only partly resolved in the *G. bulloides* record. The Gupta *et al.* (2003) analysis of *G. bulloides* reveals an intense period of upwelling between 8 and 10 ka BP that gradually decreased to a minimum of upwelling intensity ca 1500 a BP (Gupta *et al.*, 2003). Vegetation in the Horton Plains, Sri Lanka, reveals the humid character of the early Holocene, dry conditions 2 ka BP, and monsoon strengthening during the late Holocene (Premathilake and Risberg, 2003). Vegetation change in Tibet and the Himalaya also suggests an intense summer monsoon in the early Holocene (Phadtare, 2000; Hong *et al.*, 2003). The evidence is diverse and geographically widespread, leading us to hypothesise that the monsoon has been increasing in intensity since 1500 a BP. More records are needed to test this hypothesis. The records cited above, including ours, are subject to significant uncertainties in dating and proxy interpretation. In many sediment proxy records, the last millennia are more difficult to resolve because the core top sediments are soupy and easily lost or disturbed during coring. This interval requires faithful preservation of the modern (core top) environment, unusually fine sample resolution and good age control.

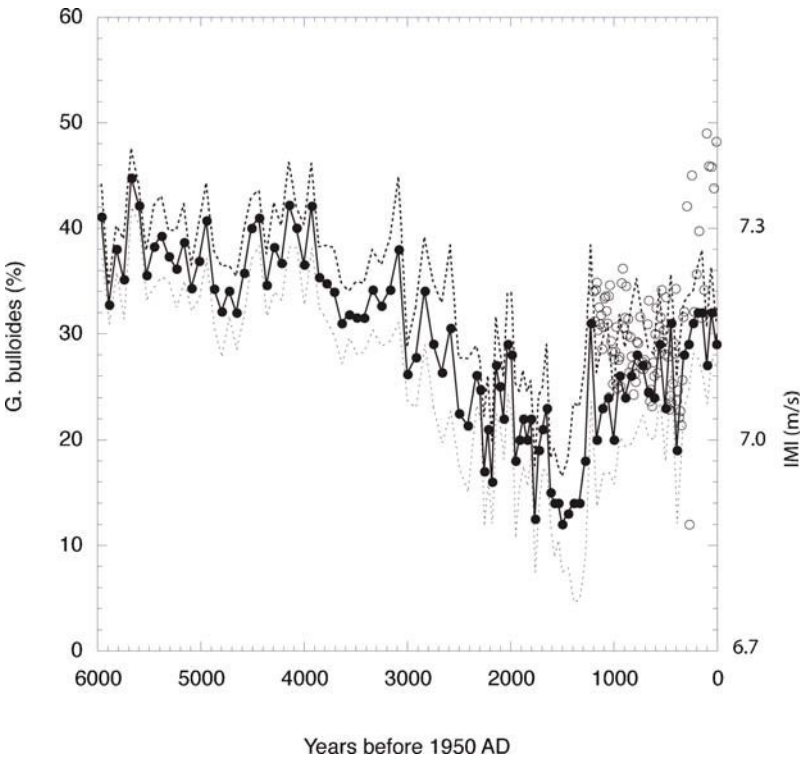

Figure 4 Time series of *G. bulloides* abundance and 95% confidence interval (dashed line) in RC2735 (filled circle) and RC2730 (open circle)

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Possible causes of the monsoon increase since 1500 a BP

At least three different influences might be responsible for the increasing trend in the monsoon since 1500 a BP. Two aspects of the solar radiation have changed. One aspect is the large slow change in summer (and winter) radiation (8% since 9 ka BP), and the second is the small (0.25%) changes in solar output inferred from radiocarbon and beryllium records. Aphelion occurs near mid July today, but 1 ka ago it occurred in mid June, and 2 ka ago it occurred during mid May. The result is that mid-May insolation has been increasing gradually since about 2 ka BP. While it is commonly assumed that July is the most significant month, if early-season radiation were important these seasonal changes could contribute to the timing of the monsoon minimum. The seasonal changes were much more gradual than the abrupt events observed in the Holocene *G. bulloides* record. Radionuclide records suggest that the solar output has also changed, although the magnitude is very small: around 0.25% (Bard *et al.*, 2000). For the last 1 ka, the timing of change in *G. bulloides* is consistent with the solar records, particularly the period of reduced *G. bulloides* abundance during the Maunder Minimum 400 a BP (Bard *et al.*, 2000). More generally, the sunspot reconstruction shows a minimum prior to the 20th century (Solanki *et al.*, 2004).

One other candidate to explain monsoon trends since 1500 a BP is agricultural and other land use changes that might have affected the monsoon. Deforestation increases albedo (reducing surface warming) while also leading to reduced latent heat flux (increasing surface warming (Brovkin *et al.*, 2006)). The history of these changes is poorly known. Some estimate that these changes became significant in the last century or two (Brovkin *et al.*, 2006; Cui *et al.*, 2007) or as early as AD 800 (Pongratz *et al.*, 2009), while others estimate that man's influence on climate began much earlier, thousands of years ago (Ruddiman, 2003). The increasing monsoon trend that we observe in recent centuries is consistent with the reduced latent heat flux influence outweighing the effect of reduced albedo. Better records are needed to test this hypothesis. The direction of future exploration regarding monsoon intensity and its forcing should work to decrease uncertainty in the proxy records and improve our understanding of the potential forcing of recent trends.

Uncertainty

To understand one source of uncertainty in *G. bulloides* abundance as a proxy for the Indian monsoon winds, we focused on quantifying the micropalaeontological counting uncertainty. Replicate counts made of all shells on the same slide by different individuals (Fig. 3) show that the difference between successive counts averages 3%. This uncertainty is attributed to the individual taxonomist's approach to identifying *G. bulloides* relative to other foraminifers. Occasionally we obtained a count that we considered erroneous because the percentage differs by a large amount from replicate counts and from adjacent samples (grey circles). Counts considered erroneous are shown on Fig. 3 but omitted from Fig. 4, which shows only the adjusted, averaged data for each core.

Another source of error arises from counting a sample of a larger population. This uncertainty relates to the probability of success in a binomial distribution. The Wald confidence interval in general use has been shown by Brown *et al.* (2001) to be flawed particularly for small samples and small and large proportions. We used the Agresti Coull Interval (Brown *et al.* (2001), where the proportion is adjusted by 'adding two successes and two failures' $CI_{AC} = p \pm 1.96 \times (p(1-p)/n)^{0.5}$ and $p = (x+2)/(n+4)$, where x is the number of *G. bulloides* counted and n is the total number of shells counted. When independent samples are taken from the same population and a confidence interval constructed, 95% of such intervals will contain the true population proportion of *G. bulloides*. Confidence intervals thus function as a measure of the precision of counting done in this study. For core RC2735, confidence interval error bars were added to the counts done by Anderson (0–99 mm). The average interval is ±5.1%. Confidence intervals could not be determined for the first part of core RC2730 (0–99 mm) because the raw counts were not preserved. The second parts (100–199 mm) of cores RC2735 and RC2730 had average confidence intervals of ±3.7% and ±3.9%, respectively. Uncertainty is lowest when the abundance of *G. bulloides* approaches 0% or 100%, and is maximal when the abundance is 50% (Table 3). The counting error could be reduced to less than ±3%, but only by counting more than 1068 shells (Table 3). The average uncertainty at the 95% level is already close to the error we attribute to taxonomist variation (3%). Furthermore, the counting error appears smaller than the difference between adjacent sediment cores from the Oman Margin (Fig. 4), and smaller than the difference we found between our initial and subsequent analysis (Core RC2730 in Fig. 2) (described in Results).

Table 3 Effect of sample size on counting uncertainty using the Agresti–Coull interval

X (# *G. bulloides*)	n (total shells)	$p = (x+2)/(n+4)$	Width: one side of the 95% confidence interval (%)
0	300	0.01	0.91
150	300	0.50	5.66
300	300	0.99	0.91
0	1068	0.00	0.26
534	1068	0.50	3.00
1068	1068	1.00	0.26

Conclusions

We improved the dating and extended our previous reconstructions of late Holocene monsoon trends using the *G. bulloides* upwelling index found in Arabian Sea sediments. In addition to the well-known decrease in the Indian summer monsoon since the early Holocene 9 ka BP, we found a minimum in the monsoon 1500 a BP, following by an increasing trend to the present. Superimposed on this increasing trend is a strong monsoon interval 1200–800 a BP and a weak monsoon interval 600–400 a BP that we correlated with the Medieval Warm Period and Little Ice Age in a previous study (Anderson *et al.*, 2002). The increasing trend is found in some but not all monsoon records from Asia and merits further study. A persistent increasing trend would have significance not only in understanding future greenhouse gas-induced change but also in understanding natural climate variability.

Acknowledgements The Radiocarbon Laboratory at the University of Arizona provided radiocarbon dates as part of their undergraduate research support program. Anne Waple and Carrie Morrill provided

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valuable comments and suggestions that improved the manuscript. The research was performed under appointment to the National Oceanic and Atmospheric Administration Ernest F. Hollings Undergraduate Scholarship Program administered by Oak Ridge Institute for Science and Education under a contract between the US Department of Energy and Oak Ridge Associated Universities.

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