Holocene aridification of India

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[1] Spanning a latitudinal range typical for deserts, the Indian peninsula is fertile instead and sustains over a billion people through monsoonal rains. Despite the strong link between climate and society, our knowledge of the long-term monsoon variability is incomplete over the Indian subcontinent. Here we reconstruct the Holocene paleoclimate in the core monsoon zone (CMZ) of the Indian peninsula using a sediment core recovered offshore from the mouth of Godavari River. Carbon isotopes of sedimentary leaf waxes provide an integrated and regionally extensive record of the flora in the CMZ and document a gradual increase in aridity-adapted vegetation from ~4,000 until 1,700 years ago followed by the persistence of aridity-adapted plants after that. The oxygen isotopic composition of planktonic foraminifer *Globigerinoides ruber* detects unprecedented high salinity events in the Bay of Bengal over the last 3,000 years, and especially after 1,700 years ago, which suggest that the CMZ aridification intensified in the late Holocene through a series of sub-millennial dry episodes. Cultural changes occurred across the Indian subcontinent as the climate became more arid after ~4,000 years. Sedentary agriculture took hold in the drying central and south India, while the urban Harappan civilization collapsed in the already arid Indus basin. The establishment of a more variable hydroclimate over the last ca. 1,700 years may have led to the rapid proliferation of water-conservation technology in south India. **Citation:** Ponton, C., L. Giosan, T. I. Eglinton, D. Q. Fuller, J. E. Johnson, P. Kumar, and T. S. Collett (2012), Holocene aridification of India, *Geophys. Res. Lett.*, *39*, L03704, doi:10.1029/2011GL050722.

1. Introduction

[2] From June to September, India receives over 80% of its annual precipitation [*Gadgil*, 2003]. The Arabian Sea branch of the monsoon delivers moisture primarily to the western Indian coast (Figure 1), where the Western Ghats range limits the penetration of rains toward the interior. The Bay of Bengal monsoon branch brings rain to most of the Indian peninsula (see auxiliary material).[1] Historical records and reconstructions spanning the last millennium show that variability in summer monsoon precipitation led to droughts, some associated with widespread famine and social disturbances, that were significantly more severe and longer lasting than anything encountered in the measured record [*Cook et al.*, 2010; *Sinha et al.*, 2011a]. The strong relationship between climate and society in India provides impetus for developing a more predictive understanding of the monsoon [*Ashfaq et al.*, 2009]. Long paleo-synoptic reconstructions would help explore the areal complexity of the monsoon, but high-resolution Holocene climate reconstructions for the Indian peninsula are conspicuously absent [*Prasad and Enzel*, 2006].

[3] Modern records from the core monsoon zone (CMZ), the region of central India that is considered representative for both the mean behavior as well as for fluctuations of the monsoon over the peninsula [*Sinha et al.*, 2011a, and references therein], show that interannual variability of summer rainfall is negatively correlated with El Niño-Southern Oscillation (ENSO) indices and strongly influenced by chaotic intraseasonal oscillations, leading to periods of increased (active) and reduced (break) precipitation [*Gadgil*, 2003; *Sinha et al.*, 2011a]. High resolution speleothem-based precipitation reconstructions in the CMZ covering the last ~1,400 years [*Sinha et al.*, 2011a, 2011b] also suggest that extended periods of break monsoon in the CMZ have been associated with most of the major droughts in the Indian subcontinent. *Sinha et al.* [2011b] argue that the immediate cause for such extensive droughts is the monsoon's persistence in a predominantly active or break mode for decades to centuries. Longer Holocene monsoon reconstructions are available from the Arabian Sea region (Figure 2). Precipitation proxy data from a stalagmite in coastal Oman shows a gradual decrease in precipitation [*Fleitmann et al.*, 2003] that is coeval with the weakening of summer monsoon winds reconstructed offshore Oman [e.g., *Gupta et al.*, 2003] and has been interpreted to reflect the southward migration of the Intertropical Convergence Zone (ITCZ) [*Fleitmann et al.*, 2007]. Although implied, it is not certain that these records can explain the hydroclimate of the Indian peninsula because of the heterogeneity of the monsoon expression at regional scales [see, e.g., *Cook et al.*, 2010; *Sinha et al.*, 2011a, 2011b].

[4] Complete Holocene monsoon reconstructions from the peninsula are limited to low resolution records and do not yield clear evidence for increased precipitation corresponding to stronger monsoon winds in the early Holocene [*Prasad and Enzel*, 2006]. Illustrating this uncertainty, an

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[1]Auxiliary materials are available at ftp://ftp.agu.org/apend/gl/2011gl050722.

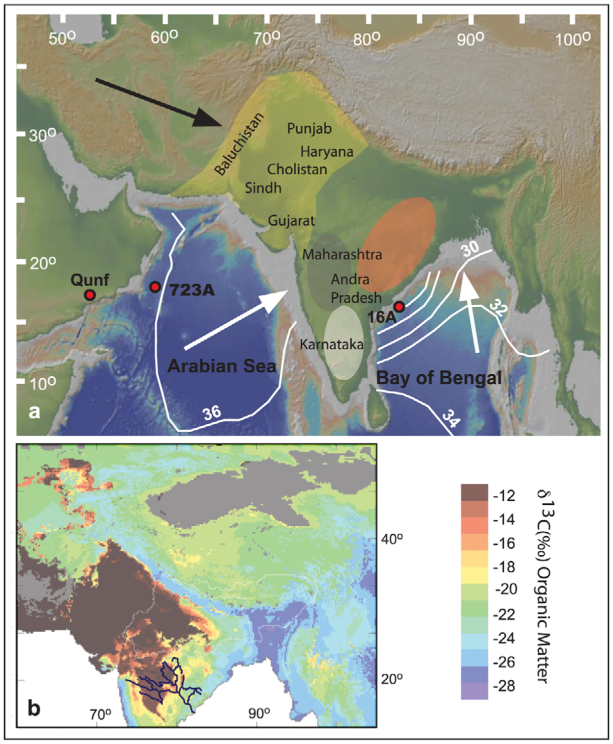

Figure 1. (a) Physiographic map of the Indian peninsula and adjacent ocean regions. Red dots show locations for core 16A, Qunf [*Fleitmann et al.*, 2003] and site 723A [*Gupta et al.*, 2003]. Black arrow is schematic for Westerly winds. White arrows indicate general directions of the Arabian Sea and Bay of Bengal branches of summer monsoon. White contours are surface salinity for June-September (in psu; see auxiliary material). Color shaded regions approximate the areal extent of past early cultures (see auxiliary material) with the corresponding modern provinces of India and Pakistan. The Indus Civilization domain is indicated by a yellow mask; Southern Neolithic is indicated by a white mask; Deccan Chalcolithic is indicated by a gray mask; and shifting cultivation domain is indicated by a red mask. (b) Average $\delta^{13}C$ of bulk terrestrial biomass in modern-day India (reprinted from *Galy et al.* [2008], with permission from Elsevier).

alternative hypothesis [*Staubwasser and Weiss*, 2006] grounded in the analysis of modern intraseasonal active-break monsoon dynamics proposes instead that the monsoon weakened only over the northernmost part of its domain over the Himalayas and their foothills, while the Indian Peninsula actually experienced an increase in monsoon intensity during the Holocene. However, an increase in precipitation is evident primarily in southern India and may reflect instead the Holocene progressive latitudinal southward shift of the summer ITCZ precipitation [*Fleitmann et al.*, 2007].

[5] To constrain monsoon variability and its effects on the Indian peninsula, we produced Holocene climate records for both the continental and oceanic realms of the Bay of Bengal monsoon branch. Terrigenous and marine components were analyzed from sediment core NGHP-01-16A (16°35.5986′N, 082°41.0070′E; 1,268 m water depth) recovered close to the Godavari River mouth, the largest non-Himalayan Indian river (see auxiliary material). The Godavari catchment (312,812 km^2, maximum elevation 920 m, Figure 1; see auxiliary material) integrates monsoon rainfall from the CMZ at the interior of the Indian peninsula and is not affected by meltwater that augments the discharge of Himalayan rivers [*Immerzeel et al.*, 2010]. We sampled hemipelagic sediments accumulating at rates higher than 0.3 m/1000 years throughout the Holocene (see auxiliary material for details on sedimentation). The cored region experiences a large seasonal range of salinity (~24–34 psu and ~2‰ change in $\delta^{18}O$ of sea water) as the monsoon freshwater plume in the Bay of Bengal disperses after the summer (see auxiliary material).

2. Methods

[6] The age model for core NGHP-16A was based on ^{14}C Accelerator Mass Spectrometry (AMS) measurements on 11 samples of mixed planktonic foraminifera (see details in auxiliary material). Stable isotope analyses of oxygen were performed using standard techniques (see auxiliary material) on planktonic foraminifera *Globigerinoides ruber* (white) at a temporal resolution of ~33 years with an analytical reproducibility better than 0.1‰ based on replicate measurements of carbonate standard NBS-19. Compound-specific carbon isotope analyses were performed on *n*-alkanoic acids (see auxiliary material) at an average sampling interval of ~220 years. Solvent-soluble

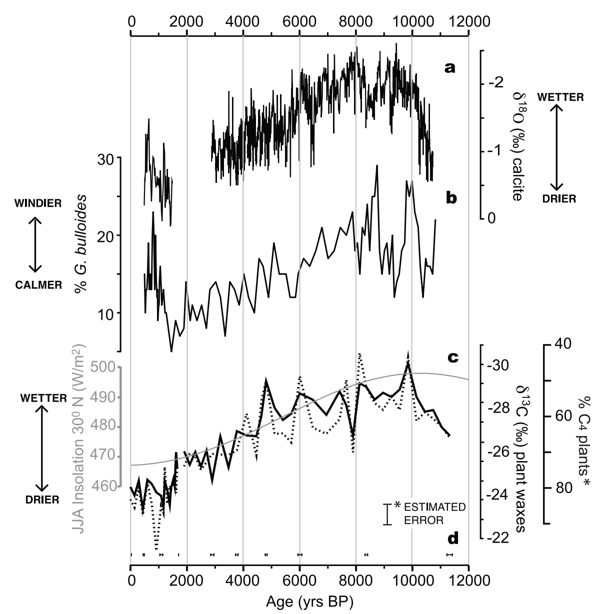

Figure 2. (a) Indian monsoon $\delta^{18}O$ record from Qunf Cave, Oman [*Fleitmann et al.*, 2003]. (b) Indian monsoon upwelling record [*Gupta et al.*, 2003]. (c) The $\delta^{13}C$ plant wax from core 16A. Black line is the weighted average of *n*-alkanoic acids C_{26}–C_{32}. Dashed line is C_{28}, the most abundant homologue in most samples. Error bar represents the maximum propagated error (1σ error; see auxiliary material) in the estimate of % C_4 plant cover (secondary axis on the far right). Grey line is the mean June-July-August insolation at 30°N [*Laskar et al.*, 2004]. (d) Calibrated radiocarbon ages (1σ error) in core 16A.

organic matter was extracted from freeze-dried sediments using a microwave accelerated reaction system. The resulting total lipid extract was saponified and the acid fraction purified and then methylated using methanol of known isotopic composition. A Gas Chromatograph with isotope ratio monitoring Mass Spectrometer (GC-irMS) was used to obtain the δ^{13}C measurements on the isolated n-alkanoic acids. All samples were analyzed in triplicate; δ^{13}C values were determined relative to a reference gas (CO_2) of known isotopic composition, introduced in pulses during each run. GC-irMS accuracy and precision are both better than 0.3‰. Results were corrected for δ^{13}C of the methyl derivative based on isotopic mass balance to derive δ^{13}C values for the original *n*-alkanoic acids (see auxiliary material).

3. Monsoon Variability in the Core Monsoon Zone

[7] The carbon isotopic composition of terrestrial plant biomass is primarily a function of the plant's specific photosynthetic pathway and isotopic composition of atmospheric CO_2 [*Farquhar et al.*, 1989], with environmental conditions exerting a minimal influence. These isotopic signatures also manifest themselves in vascular plant epicuticular wax lipids [e.g., *Tipple and Pagani*, 2010]. Leaf wax δ^{13}C records (i.e., for C_{26} to C_{32} *n*-alkanoic acids; hereafter $\delta^{13}C_{wax}$; see auxiliary material) have been used extensively to reconstruct past changes in the balance of C_3 vs. C_4 vegetation [*Feakins et al.*, 2005; *Eglinton and Eglinton*, 2008]. C_4 vegetation is favored by aridity, high temperature, and low atmospheric CO_2 conditions over C_3 plants. Given the minimal variability in annual sea surface temperature in the northern Indian Ocean region [*Govil and Naidu*, 2010; *Anand et al.*, 2008; *Rashid et al.*, 2007; *Govil and Naidu*, 2011] and increasing CO_2 over the Holocene, our leaf wax δ^{13}C record reflects the integrated rainfall variations or aridity in the Godavari river catchment where natural vegetation cover is a mixture of savanna, tropical grassland, and tropical forest [*Asouti and Fuller*, 2008]. Modeled bulk organic carbon δ^{13}C values corresponding to this modern biome mixture vary from ca. −12 to −26‰ (Figure 1) [*Galy et al.*, 2008]. In our core, $\delta^{13}C_{wax}$ (Figure 2) exhibits a large range of variation (∼−23‰ to ∼−30‰).

[8] Based on prior δ^{13}C measurements of *n*-alkanoic acids isolated from different plant species [*Chikaraishi et al.*, 2004], we calculated isotopic end members of −37.7 ± 1.8‰ and −21.1 ± 1.4‰ for C_3 and C_4 plants (see auxiliary material), respectively. Using a simple isotopic mass balance we estimate that the proportion of C_4 vegetation cover in central India increased from approximately 50% to more than 75% during the Holocene (Figure 2). The average error in the changes in abundance of C_3 vs. C_4 plants (see auxiliary material) based on the *Chikaraishi et al.* [2004] species survey amounts to ±6.3%, uncertainty which is considerably less than the changes in our record (Figure 2). Although there is significant carbon isotopic variability within C_4 and especially C_3 plants and their corresponding waxes [*Freeman and Collarusso*, 2001; *Chikaraishi et al.*, 2004; *Tipple and Pagani*, 2010], which limits the ability to place tight constraints on changes in the proportion of C_4 plants, the magnitude of change during the Holocene reflects a significant shift in vegetation type. The range of variation for δ^{13}C composition among individual C_{26} to C_{32} *n*-alkanoic acid homologues also decreased from early to late Holocene (see auxiliary material). We speculate that this trend is also a response to increased aridity, and may reflect the narrower range of δ^{13}C values expressed by C_4 plants [*Freeman and Collarusso*, 2001] or a reduction in plant diversity [*Rommerskirchen et al.*, 2003]. Crassulacean acid metabolism (CAM) plants, which can utilize both C_3 and C_4 carbon fixation pathways and have an intermediate δ^{13}C range, are common in central India [*Asouti and Fuller*, 2008] and may lead to an underestimation of C_4 cover, but also represent aridity-adapted vegetation. Additionally, anthropogenic contributions to the C_4 signal through cultivation cannot be completely discounted but are likely to have been small until the 19th century when massive and permanent deforestation of the Eastern Ghats took place [*Hill*, 2008]. Prior to this large scale deforestation, the shifting cultivation style typical for the Eastern Ghats (Figure 1a) where most C_3 flora occurs in the Godavari watershed, did not favor large changes in C_3 vs. C_4 plants (see auxiliary material). Early farming in the Deccan (Figure 1a) replaced C_4-dominated savannah and adjacent woodland with C_4 cultivated plants, but also affected the Western Ghats C_3 forests, which comprise only a small part of the Godavari's headwaters. Rice, a C_3 plant that was cultivated in coastal regions after 3,000 years ago, would have ameliorated rather than accentuated the trend toward C_4 flora dominance in the late Holocene.

[9] The inferred change in vegetation structure is comparable in magnitude to a major glacial to interglacial ecosystem alteration (cf., ∼20% shift toward more C_4 plants in the Himalayas from the Last Glacial Maximum to early Holocene [*Galy et al.*, 2008]). Our own $\delta^{13}C_{wax}$ measurements on glacial-age samples at ∼26 ka BP show ^{13}C-enriched values (−22.3‰; see auxiliary material), implying that the vegetation cover in the Godavari catchment was similarly populated with C_4 plants (>85%). After a humid early Holocene when the proportion of C_4 plants oscillated significantly, there was a marked change towards more positive $\delta^{13}C_{wax}$ values that persisted until ca. 1,700 years ago, reflecting the increasing aridification of central India. This increase in aridity is most evident after 4,000 BP (Figure 2) when the δ^{13}C values for all C_{26} to C_{32} *n*-alkanoic acid homologues shift to values beyond their previous range of variability. The last ∼1,700 years appear to be anomalously arid with an apparent dominance of C_4 vegetation. The Holocene aridification of central India supports the view that changes in the seasonality of Northern Hemisphere insolation associated with the orbital precession, led to progressively weaker monsoons [*Fleitmann et al.*, 2007]. In concert with previous reconstructions in the Arabian Sea region [*Fleitmann et al.*, 2003; *Gupta et al.*, 2003] and northern Bay of Bengal [*Kudrass et al.*, 2001], this aridification of the core monsoon zone shows that the Indian monsoon displayed a largely coherent response during the Holocene.

[10] We explored further the changes in aridity for the last 4,500 years by examining the oxygen isotope composition of planktonic foraminifer *Globigerinoides ruber* ($\delta^{18}O_{ruber}$) from core NGHP-01-16A (Figure 3). After applying a positive correction for the effects of post-glacial ice-sheet decay varying between 0 and 0.07‰ (see auxiliary material), $\delta^{18}O_{ruber}$ should record surface water conditions in the Bay of Bengal. The relatively low δ^{18}O

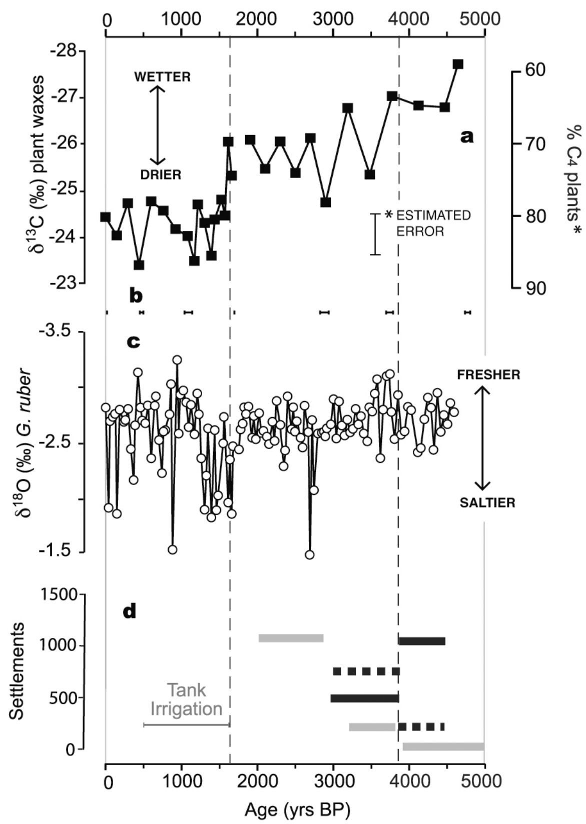

Figure 3. (a) δ^{13}C plant wax record from core 16A as the weighted average of *n*-alkanoic acids C_{26}–C_{32}. Error bar represents the maximum propagated error (1σ error; see auxiliary material) in the estimate of % C_4 plant cover. Vertical black dashed lines identify steps in the aridification at ca. 4,000 and 1,700 years BP. (b) Calibrated radiocarbon ages (1σ error) in core 16A. (c) δ^{18}O measured on *Globigerinoides ruber* from core 16A; values are corrected for ice volume effects. (d) Number of settlements based on archaeological data expressed as totals over culturally defined time intervals (see auxiliary material). In solid gray, sites from the Deccan Plateau (Andhra Pradesh, Karnataka, Maharashtra). In solid black, Indus (Harappan) sites from the dry Baluchistan, Sindh, Gujarat, Cholistan and lower Punjab. In dashed black, sites from rainier upper Punjab and Haryana. The drought-prone regime in the late Holocene (after 1,700 years BP) coincides with the flourishing of water tank construction.

values of our record and the lack of a clear trend in the δ^{18}O time series is not surprising for this region reflecting large fluvial discharge and the preformed waters of these fluvial sources [*Breitenbach et al.*, 2010]. Freshwater from Godavari, augmented by several other large rivers, together with direct precipitation over the Bay and the water exchange with adjacent regions of the Indian Ocean act to buffer local variability [*Schott and McCreary*, 2001] (see auxiliary material). Relatively stable sea surface conditions characterize the interval between ~4,500 and 3,000 years BP (Figure 3) with variability between ~−2.3‰ and −3.3‰. However, after 3,000 years BP, and especially over the last 1,700 years, $\delta^{18}O_{ruber}$ values vary between ~−1.4‰ and −3.3‰ exhibiting marked positive excursions. Similar to other tropical regions, Holocene sea surface temperature (SST) fluctuations in the northern Indian Ocean were small (up to 2.5°C) after ~8,000 years BP (see *Anand et al.* [2008] and *Govil and Naidu* [2010] for eastern Arabian Sea, *Rashid et al.* [2007] for the Andaman Sea, and *Govil and Naidu* [2011] for the western Bay of Bengal) accounting for a maximum of 0.56‰ change in $\delta^{18}O_{ruber}$ (see auxiliary material). Thirteen excursions of sub-millennial duration occurring in our late Holocene record are beyond the variance that can be explained by temperature variations. Simulations with a fully coupled atmosphere-ocean global climate model [*LeGrande and Schmidt*, 2009] also suggest that changes in precipitation sources over the Bay of Bengal were minimal in the last 6,000 years compared to earlier in the Holocene.

[11] In this context, we interpret the positive $\delta^{18}O_{ruber}$ excursions after 3,000 years BP to reflect increased salinity events in the Bay of Bengal during drier intervals reminiscent of the extended droughts documented for the last millennium [*Cook et al.*, 2010; *Sinha et al.*, 2011a]. The lack of a corresponding increase in variance in the $\delta^{13}C_{wax}$ record over the late Holocene may reflect the buffering of short term terrestrial sedimentary signals within the Godavari watershed and/or sluggish recovery of C_3 continental flora in a variable hydroclimatic regime. Records from cores with lower sedimentation rates from the northern Bay of Bengal and the Andaman Sea as well as eastern Arabian Sea also argue for higher salinities in the late Holocene [*Kudrass et al.*, 2001; *Rashid et al.*, 2007; *Govil and Naidu*, 2010]. The $\delta^{18}O_{ruber}$ excursions are particularly prominent between ~1,700 and 1,300 years BP, coincident with the Holocene monsoon minimum in the wind proxy reconstruction in the Arabian Sea [*Anderson et al.*, 2010] (Figure 3). Considering that the background $\delta^{18}O_{ruber}$ over the last 1,300 years indicates surface salinities as low as in the middle Holocene or lower, it is reasonable to assume that the levels of precipitation were high outside these dry episodes, consistent with the increase in monsoon winds in the Arabian Sea during the same interval [*Anderson et al.*, 2010]. In the eastern Arabian Sea, small annual mean SST variability (less than 1°C) reconstructed on *G. ruber* contrasts with the increased seasonality indicated by coeval temperature record on *G. bulloides* over the past ~4,000 years [*Anand et al.*, 2008], lending further support to the idea that $\delta^{18}O_{ruber}$ excursions primarily reflect changes in monsoon variability. Recent foraminifer-based records of *Govil and Naidu* [2011] and *Chauhan et al.* [2010] suggests increased monsoon variability after ca. 3,500 and 2,200 years BP respectively whereas a speleothem record from NE India [*Adkins et al.*, 2011; *Breitenbach*, 2010] shows increased monsoon variance in the last ~2,000 years. Taken together, these reconstructions suggest that monsoon variability increased coherently over the Indian peninsula in the late Holocene, although at sub-millennial timescale the variability may have been anti-phased between the peninsula and northeastern India [*Sinha et al.*, 2011b].

[12] A recent reconstruction of the Australian-Indonesian monsoon shows a relatively dry early to middle Holocene and an increase in precipitation during the past ~2,500 years [*Mohtadi et al.*, 2011]. The out-of-phase behavior of rainfall in central India in the Northern Hemisphere versus Indonesia in the Southern Hemisphere is in agreement with a primary orbital control on the monsoon in the Indian

Ocean. Our terrestrial and oceanic records further suggest that aridification intensified in the CMZ in the late Holocene through a series of sub-millennial dry episodes. We note that these episodes occurred preferentially in late Holocene when the intensity of Northern Hemisphere summer insolation reached its minimum. El Niño frequency/intensity also increased in late Holocene [e.g., *Moy et al.*, 2002]. However, understanding how the CMZ dry episodes relate to changes in Northern Hemisphere climatic state or to coupled ocean-atmosphere zonal modes, such as El Niño–Southern Oscillation, would require complete Holocene climate reconstructions from this region at subdecadal resolution.

4. Aridification and Cultural Change

[13] The dramatic effects of monsoon variability on the well-being of the peninsular population over the historical period are well-documented [*Cook et al.*, 2010]. Regardless of the exact mechanisms leading to aridification of central India during the Holocene, our new data suggest that this symbiotic human-monsoon relationship may have existed since prehistory. The significant aridification recorded after ca. 4,000 years ago may have spurred the widespread adoption of sedentary agriculture in central and south India capable of providing surplus food in a less secure hydroclimate [*Asouti and Fuller*, 2008]. Archaeological site numbers and the summed probability distributions of calibrated radiocarbon dates from archaeological sites, which serve as proxies of agricultural population, increase markedly after 4,000 BP in peninsular India (Figure 3; see auxiliary material) for cultures of Southern India and the Deccan Plateau. In contrast, the same process of drying elicited the opposite response (Figure 3) in the already arid northwestern region of the subcontinent along the Indus River (Figure 1a). From $\sim$3,900 to 3,200 years BP, the urban Harappan civilization entered a phase of protracted collapse. Late Harrapan rural settlements became instead more numerous in the rainier regions at the foothills of the Himalaya and in the Ganges watershed [*Madella and Fuller*, 2006]. During the Iron Age, after ca. 3,200 years BP, adaptation to semi-arid conditions in central and south India appears to have been well established with $\sim$60% of sites in areas with <1000 mm of rainfall today and a significant number of sites (18%) in areas with <600 mm (see auxiliary material). Later, the rapid increase in rainwater harvesting structures that occurred after 1,700 BP in the semi-arid regions of south India [*Gunnell et al.*, 2007] points to an expansion of the cultural adaptation to an additional increase in aridity (Figure 3). Although tentative, these correlations between hydroclimate and cultural changes in the Indian subcontinent suggest distinct societal responses to climate stress and underline the importance of understanding the history of the monsoon in a dynamic context at synoptic scales both for interpreting the past, but also for providing the long-term context for the short instrumentally observed variability.

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References

Adkins, J. F., S. Breitenbach, B. Plessen, N. Marwan, D. C. Lund, P. Huybers, and G. H. Haug (2011), Holocene history of δ^{18}O and grayscale from a stalagmite in NE India, with implications for monsoon and ENSO variability, Abstract PP22B-06, presented at 2011 Fall Meeting, AGU, San Francisco, Calif., 5–9 Dec.

Anand, P., D. Kroon, A. D. Singh, R. S. Ganeshram, G. Ganssen, and H. Elderfield (2008), Coupled sea surface temperature–seawater δ^{18}O reconstructions in the Arabian Sea at the millennial scale for the last 35 ka, *Paleoceanography*, *23*, PA4207, doi:10.1029/2007PA001564.

Anderson, D. M., C. K. Baulcomb, A. K. DuVivier, and A. K. Gupta (2010), Indian summer monsoon during the last two millennia, *J. Quat. Sci.*, *25*(6), 911–917, doi:10.1002/jqs.1369.

Ashfaq, M., Y. Shi, W. W. Tung, R. J. Trapp, X. J. Gao, J. S. Pal, and N. S. Diffenbaugh (2009), Suppression of south Asian summer monsoon precipitation in the 21st century, *Geophys. Res. Lett.*, *36*, L01704, doi:10.1029/2008GL036500.

Asouti, E., and D. Q. Fuller (2008), *Trees and Woodlands in South India. Archaeological Perspectives*, Left Coast, Walnut Creek, Calif.

Breitenbach, S. (2010), Changes in monsoonal precipitation and atmospheric circulation during the Holocene reconstructed from stalagmites from Northeastern India, *Sci. Tech. Rep. 10/06*, Dtsch. GeoForschungsZent., Potsdam, Germany. doi:10.2312/GFZ.b103-10060.

Breitenbach, S. F. M., J. F. Adkins, H. Meyer, N. Marwan, K. K. Kumar, and G. H. Haug (2010), Strong influence of water vapor source dynamics on stable isotopes in precipitation observed in southern Meghalaya, *Earth Planet. Sci. Lett.*, *292*(1–2), 212–220, doi:10.1016/j.epsl.2010.01.038.

Chauhan, O. S., E. Vogelsang, N. Basavaiah, and U. S. A. Kader (2010), Reconstruction of the variability of the southwest monsoon during the past 3 ka, from the continental margin of the southeastern Arabian Sea, *J. Quat. Sci.*, *25*, 798–807, doi:10.1002/jqs.1359.

Chikaraishi, Y., H. Naraoka, and S. R. Poulson (2004), Hydrogen and carbon isotopic fractionations of lipid biosynthesis among terrestrial (C3, C4 and CAM) and aquatic plants, *Phytochemistry*, *65*(10), 1369–1381, doi:10.1016/j.phytochem.2004.03.036.

Cook, E. R., K. J. Anchukaitis, B. M. Buckley, R. D. D'Arrigo, G. C. Jacoby, and W. E. Wright (2010), Asian monsoon failure and megadrought during the last millennium, *Science*, *328*(5977), 486–489, doi:10.1126/science.1185188.

Eglinton, T. I., and G. Eglinton (2008), Molecular proxies for paleoclimatology, *Earth Planet. Sci. Lett.*, *275*, 1–16, doi:10.1016/j.epsl.2008.07.012.

Farquhar, G. D., J. R. Ehleringer, and K. T. Hubick (1989), Carbon isotope discrimination and photosynthesis, *Annu. Rev. Plant Physiol. Mol. Biol.*, *40*, 503–537, doi:10.1146/annurev.pp.40.060189.002443.

Feakins, S. J., P. B. deMenocal, and T. I. Eglinton (2005), Biomarker records of late Neogene changes in northeast African vegetation, *Geology*, *33*(12), 977–980, doi:10.1130/G21814.1.

Fleitmann, D., S. J. Burns, M. Mudelsee, U. Neff, J. Kramers, A. Mangini, and A. Matter (2003), Holocene forcing of the Indian monsoon recorded in a stalagmite from southern Oman, *Science*, *300*(5626), 1737–1739, doi:10.1126/science.1083130.

Fleitmann, D., et al. (2007), Holocene ITCZ and Indian monsoon dynamics recorded in stalagmites from Oman and Yemen, *Quat. Sci. Rev.*, *26*(1–2), 170–188, doi:10.1016/j.quascirev.2006.04.012.

Freeman, K. H., and L. A. Collarusso (2001), Molecular and isotopic records of C4 grassland expansion in the late Miocene, *Geochim. Cosmochim. Acta*, *65*, 1439–1454, doi:10.1016/S0016-7037(00)00573-1.

Gadgil, S. (2003), The Indian monsoon and its variability, *Annu. Rev. Earth Planet. Sci.*, *31*, 429–467, doi:10.1146/annurev.earth.31.100901.141251.

Galy, V., L. Francois, C. France-Lanord, P. Faure, H. Kudrass, F. Palhol, and S. K. Singh (2008), C4 plants decline in the Himalayan basin since the Last Glacial Maximum, *Quat. Sci. Rev.*, *27*(13–14), 1396–1409, doi:10.1016/j.quascirev.2008.04.005.

Govil, P., and P. D. Naidu (2010), Evaporation-precipitation changes in the eastern Arabian Sea for the last 68 ka: Implications on monsoon variability, *Paleoceanography*, *25*, PA1210, doi:10.1029/2008PA001687.

Govil, P., and P. D. Naidu (2011), Variations of Indian monsoon precipitation during the last 32 kyr reflected in the surface hydrography of the western Bay of Bengal, *Quat. Sci. Rev.*, *30*, 3871–3879, doi:10.1016/j.quascirev.2011.10.004.

Gunnell, Y., K. Anupama, and B. Sultan (2007), Response of the south Indian runoff-harvesting civilization to northeast monsoon rainfall variability during the last 2000 years: instrumental records and indirect evidence, *Holocene*, *17*(2), 207–215, doi:10.1177/0959683607075835.

Gupta, A. K., D. M. Anderson, and J. T. Overpeck (2003), Abrupt changes in the Asian southwest monsoon during the Holocene and their links to the North Atlantic Ocean, *Nature*, *421*(6921), 354–357, doi:10.1038/nature01340.

Hill, C. V. (2008), *South Asia: An Environmental History*, 329 pp., ABC-CLIO, Santa Barbara, Calif.

Immerzeel, W. W., L. P. H. van Beek, and M. F. P. Bierkens (2010), Climate change will affect the Asian water towers, *Science*, *328*, 1382–1385, doi:10.1126/science.1183188.

Kudrass, H. R., A. Hofmann, H. Doose, K. Emeis, and H. Erlenkeuser (2001), Modulation and amplification of climatic changes in the Northern Hemisphere by the Indian summer monsoon during the past 80 k.y., *Geology*, *29*(1), 63–66, doi:10.1130/0091-7613(2001)029<0063:MAAOCC>2.0.CO;2.

Laskar, J., P. Robutel, F. Joutel, M. Gastineau, A. C. M. Correia, and B. Levrard (2004), A long-term numerical solution for the insolation quantities of the Earth, *Astron. Astrophys.*, *428*(1), 261–285, doi:10.1051/0004-6361:20041335.

LeGrande, A. N., and G. A. Schmidt (2009), Sources of Holocene variability of oxygen isotopes in paleoclimate archives, *Clim. Past*, *5*(3), 441–455, doi:10.5194/cp-5-441-2009.

Madella, M., and D. Q. Fuller (2006), Palaeoecology and the Harappan civilisation of South Asia: A reconsideration, *Quat. Sci. Rev.*, *25*(11–12), 1283–1301, doi:10.1016/j.quascirev.2005.10.012.

Mohtadi, M., D. W. Oppo, S. Steinke, J.-B. W. Stuut, R. De Pol-Holz, D. Hebbeln, and A. Luckge (2011), Glacial to Holocene swings of the Australian-Indonesian monsoon, *Nat. Geosci.*, *4*, 540–544, doi:10.1038/ngeo1209.

Moy, C. M., G. O. Seltzer, D. T. Rodbell, and D. M. Anderson (2002), Variability of El Nino/Southern Oscillation activity at millennial timescales during the Holocene epoch, *Nature*, *420*(6912), 162–165, doi:10.1038/nature01194.

Prasad, S., and Y. Enzel (2006), Holocene paleoclimates of India, *Quat. Res.*, *66*(3), 442–453, doi:10.1016/j.yqres.2006.05.008.

Rashid, H., B. P. Flower, R. Z. Poore, and T. M. Quinn (2007), A similar to 25 ka Indian Ocean monsoon variability record from the Andaman Sea, *Quat. Sci. Rev.*, *26*(19–21), 2586–2597, doi:10.1016/j.quascirev.2007.07.002.

Rommerskirchen, F., G. Eglinton, L. Dupont, U. Guntner, C. Wenzel, and J. Rullkotter (2003), A north to south transect of Holocene southeast Atlantic continental margin sediments: Relationship between aerosol transport and compound-specific δ^{13}C land plant biomarker and pollen records, *Geochem. Geophys. Geosyst.*, *4*(12), 1101, doi:10.1029/2003GC000541.

Schott, F. A., and J. P. McCreary (2001), The monsoon circulation of the Indian Ocean, *Prog. Oceanogr.*, *51*(1), 1–123, doi:10.1016/S0079-6611(01)00083-0.

Sinha, A., M. Berkelhammer, L. Stott, M. Mudelsee, H. Cheng, and J. Biswas (2011a), The leading mode of Indian summer monsoon precipitation variability during the last millennium, *Geophys. Res. Lett.*, *38*, L15703, doi:10.1029/2011GL047713.

Sinha, A., L. Stott, M. Berkelhammer, H. Cheng, R. L. Edwards, B. Buckley, M. Aldenderfer, and M. Mudelsee (2011b), A global context for megadroughts in monsoon Asia during the past millennium, *Quat. Sci. Rev.*, *30*(1–2), 47–62, doi:10.1016/j.quascirev.2010.10.005.

Staubwasser, M., and H. Weiss (2006), Holocene climate and cultural evolution in late prehistoric-early historic west Asia—Introduction, *Quat. Res.*, *66*(3), 372–387, doi:10.1016/j.yqres.2006.09.001.

Tipple, B. J., and M. Pagani (2010), A 35 Myr North American leaf-wax compound-specific carbon and hydrogen isotope record: Implications for C4 grasslands and hydrologic cycle dynamics, *Earth Planet. Sci. Lett.*, *299*, 250–262, doi:10.1016/j.epsl.2010.09.006.

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