Holocene aridification of India

Camilo Ponton,1,2 Liviu Giosan,2 Tim I. Eglinton,3,4 Dorian Q. Fuller,5 Joel E. Johnson,6 Pushpendra Kumar,7 and Tim S. Collett8

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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.


1. Introduction

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].

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].

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

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 (see auxiliary material for details on sedimentation). The cored region experiences a large seasonal range of salinity (~24–34 psu and ~2‰ change in δ18O 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 14C 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 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 δ13C of bulk terrestrial biomass in modern-day India (reprinted from Galy et al. [2008], with permission from Elsevier).

Figure 2. (a) Indian monsoon δ18O record from Qunf Cave, Oman [Fleitmann et al., 2003]. (b) Indian monsoon upwelling record [Gupta et al., 2003]. (c) The δ13C plant wax from core 16A. Black line is the weighted average of n-alkanoic acids C26–C32. Dashed line is C28, the most abundant homologue in most samples. Error bar represents the maximum propagated error (1σ error; see auxiliary material) in the estimate of % C4 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 $\delta^{13}C$ measurements on the isolated n-alkanoic acids. All samples were analyzed in triplicate; $\delta^{13}C$ 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 $\delta^{13}C$ of the methyl derivative based on isotopic mass balance to derive $\delta^{13}C$ 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 $\delta^{13}C$ records (i.e., for C$_{26}$ to C$_{32}$ n-alkanoic acids; hereafter $\delta^{13}C_{\text{wax}}$; 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 $\delta^{13}C$ 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 $\delta^{13}C$ values corresponding to this modern biome mixture vary from ca. −12 to −26‰ (Figure 1) [Galy et al., 2008]. In our core, $\delta^{13}C_{\text{wax}}$ (Figure 2) exhibits a large range of variation (~23% to ~30%).

[8] Based on prior $\delta^{13}C$ 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 Collarussso, 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 $\delta^{13}C$ 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 $\delta^{13}C$ values expressed by C$_4$ plants [Freeman and Collarussso, 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 $\delta^{13}C$ 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$_3$ 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 $\delta^{13}C_{\text{wax}}$ measurements on glacial-age samples at ~26 ka BP show C$_4$-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 $\delta^{13}C_{\text{wax}}$ 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 $\delta^{13}C$ 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$_3$ 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 ($\delta^{18}O_{\text{ruber}}$) 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), $\delta^{18}O_{\text{ruber}}$ should record surface water conditions in the Bay of Bengal. The relatively low $\delta^{18}O$...
over the last 1,700 years, fluvial sources ([C0]large fluvial discharge and the preformed waters of these [C24]/C0 BP (Figure 3) with variability between buffer local variability ([C0], 2001] (see Breitenbach et al., 2001; Rashid et al., 2007; Govil and Naidu, 2010). The [C24]/C0 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 [C24]/C0 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 [C24]/C0 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].

Figure 3. (a) $\delta^{13}$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) $\delta^{18}$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, Gujrat, 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 $\delta^{18}$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 0.5‰ change in $\delta^{18}$O$_{ruber}$ (see auxiliary material). Thirteen excursions of sub-millenial 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].

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 [Mohrati 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., Moey 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 sub-decadal 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 ~3,900 to 3,200 years BP, the urban Harappan civilization entered a phase of protracted collapse. Late Harappan rural settlements became instead urban Harappan civilization [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


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


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


