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Short communication: Massive erosion in monsoonal central India linked to late Holocene land cover degradation

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Abstract. Soil erosion plays a crucial role in transferring sediment and carbon from land to sea, yet little is known about the rhythm and rates of soil erosion prior to the most recent few centuries. Here we reconstruct a Holocene erosional history from central India, as integrated by the Godavari River in a sediment core from the Bay of Bengal. We quantify terrigenous fluxes, fingerprint sources for the lithogenic fraction and assess the age of the exported terrigenous carbon. Taken together, our data show that the monsoon decline in the late Holocene significantly increased soil erosion and the age of exported organic carbon. This acceleration of natural erosion was later exacerbated by the Neolithic adoption and Iron Age extensification of agriculture on the Deccan Plateau. Despite a constantly elevated sea level since the middle Holocene, this erosion acceleration led to a rapid growth of the continental margin. We conclude that in monsoon conditions aridity boosts rather than suppresses sediment and carbon export, acting as a monsoon erosional pump modulated by land cover conditions.

1 Soil erosion in the Holocene

On decadal to millennial timescales, climate is the principal natural control on soil erosion via changes in temperature and precipitation as well as their impact on vegetation type and cover (Allen and Breshears, 1998; Reichstein et al., 2013). Global sediment budgets for the Holocene indicate that humans surpassed these natural controls and became the main driver of soil erosion by at least 2000 years ago (Montgomery, 2007; Wilkinson and McElroy, 2007; Dotterweich, 2013). Transfer of sediment, carbon and solutes from land to ocean is of crucial importance for understanding continental margin architecture as well as carbon and other elemental cycles. For example, soils contain about 2 times more carbon than the atmosphere and, as a result, small changes in the residence time of organic carbon in soils can significantly affect the atmospheric inventory of carbon dioxide (Lal, 2004). Besides heterotrophic microbial respiration, erosion is the principal process that releases carbon from soils. Eroded carbon can subsequently be degraded/reburied along the aquatic continuum to the ocean (Stallard, 1998; Aufdenkampe et al., 2011; van Oost et al., 2012).

In the absence of historical documentation of human impacts, the complexity of soil erosion hampers the reconstruction of carbon transfer processes prior to the last few centuries (e.g., Hoffmann et al., 2013; Dotterweich, 2013; Vanwalleghem et al., 2017). Consequently, global carbon budgets implicitly assume steady state conditions for lateral transport and carbon degradation along the aquatic continuum in pre-industrial times (Battin et al., 2009; Regnier et al., 2013; Chappell et al., 2016). In contrast, abundant archaeological and geological evidence (e.g., van Andel et al., 1990; Bork and Lang, 2003; Bayon et al., 2012; Dotterweich, 2013) as well as modeling (Kaplan et al., 2010; Wang et al., 2017) suggests widespread impacts of early human land use on continental landscapes, soil erosion and associated carbon transfer processes.

Here we present a soil erosion history from the Indian Peninsula recorded in a sediment core retrieved near the mouth of the Godavari River (Fig. 1) in the Bay of Bengal (NGHP-01-16A at 16° 35.6′ N, 82°41.0′ E; 1268 m water depth; Collett et al., 2015). The age model for the core based on 11 radiocarbon dates on mixed planktonic foraminifera was previously published by Ponton et al. (2012). The Godavari Basin was not affected by tectonics on the Holocene timescale or by glacial/snow meltwater and strong orographic precipitation, which augment and complicate the water and sediment discharge of the larger Himalayan rivers like the Ganges or Brahmaputra. Instead, it integrates rainfall from the Core Monsoon Zone (CMZ), the region of central India that is representative of both the mean monsoon regime and its fluctuations over the peninsula (see Ponton et al., 2012, and references therein). Consequently, over 90 % of the Godavari's water discharge into the Bay of Bengal derives from summer monsoon precipitation (Rao et al., 2005), making its sedimentary deposits a prime target for continental climate reconstructions and a repository for sedimentary proxies of erosion prior and after the Neolithic adoption of agriculture in central India.

2 The Godavari sediment system

Originating at an elevation of 920 m in the Sahyadri coastal range (aka Western Ghats) near the Arabian Sea coast, the Godavari crosses the entire Indian Peninsula toward the Bay of Bengal (Fig. 1a). Currently the water discharge of the river is $\sim 85\,\mathrm{km^3\,yr^{-1}}$ with a sediment load of $\sim 175\,\mathrm{Mt\,yr^{-1}}$ (Syvitski and Saito, 2007). Because the coastal range limits penetration of the Arabian Sea moisture delivered by the monsoon, precipitation in the Godavari Basin primarily originates from the Bay of Bengal (Gunnell et al., 2007). As a result, the climate is most humid at the coast (i.e., Eastern

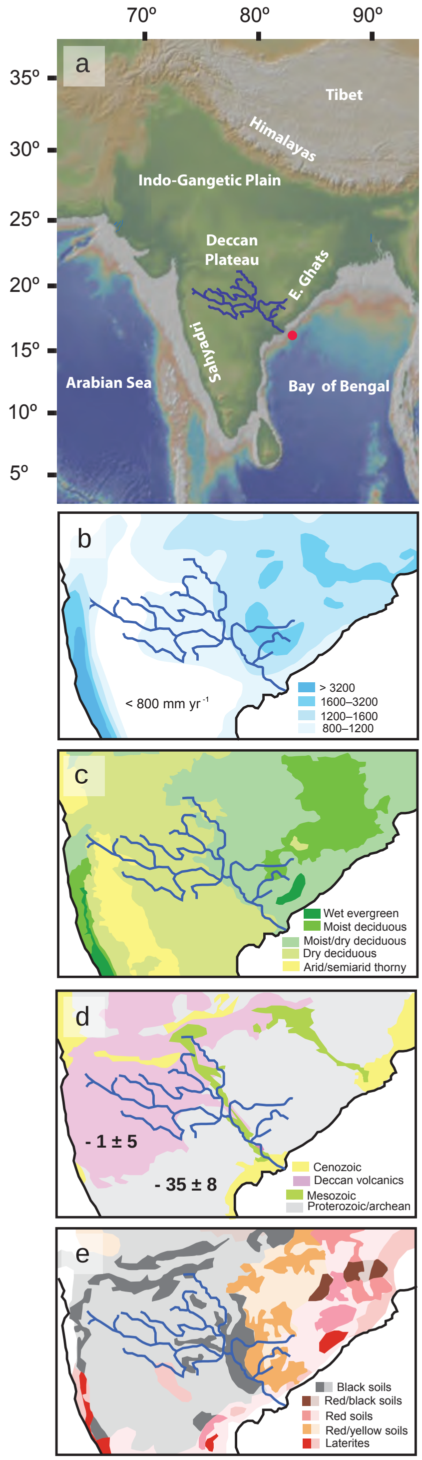

Figure 1. Godavari River drainage basin in its **(a)** physiographical, **(b)** hydroclimatic (Asouti and Fuller, 2008), **(c)** ecological (Asouti and Fuller, 2008), **(d)** geological (Bikshamaiah and Subramanian, 1988) and **(e)** soil cover (NBSS & LUP, 1983) context. Core NGHP-01-16A location is indicated in **(a)** by the red dot. Average bedrock εNd values are shown in **(d)**.

Ghats range) and becomes increasingly arid toward the interior on the Deccan Plateau (Fig. 1b). The natural vegetation reflects this gradual decrease in moisture: the headwaters on the Deccan Plateau are dominated by C_4-plant thornbush savannah adapted to dry conditions, whereas C_3 flora (deciduous forests) are dominant in the Eastern Ghats (Asouti and Fuller, 2008; Fig. 1c).

Sediments transported by the Godavari are sourced from two major geological units (Bikshamaiah and Subramanian, 1988). The upper river basin developed on the Deccan Traps, a large igneous province consisting of relatively young flood basalts (Cretaceous to early Neogene) that largely span the Deccan Plateau. The lower river basin developed over old Proterozoic to Archaean crystalline igneous/metamorphic rocks of the Indian Craton (Fig. 1d). The relatively young Deccan basalts retain a highly radiogenic mantle-derived Nd isotope composition (εNd of $+1 \pm 5$), while the old continental crust of the Indian Craton has a relatively unradiogenic isotopic composition (εNd of -35 ± 8), yielding a sharp contrast between geological end-members. Thus, the sediment provenance for the Godavari sediments can be deduced from the Nd isotopic signatures of the detrital inorganic fraction in our core because the Nd signal remains unmodified through bedrock weathering processes (McLennan and Hemming, 1992; DePaolo, 1988).

Black soils cover the Deccan Plateau, whereas red soils are generally typical for the Eastern Ghats (Bhattacharyya et al., 2003; Fig. 1d). Although both types of soils have been affected by land use since prehistorical times, the black soils of the arid to semiarid Deccan Plateau appear to be the most degraded at present (Singh et al., 1992). Intense erosion within the basin is reflected by the inordinately large sediment load of the Godavari (Bikshamaiah and Subramanian, 1988) similar to other monsoonal rivers (Summerfield and Hulton, 1994). In contrast to the dynamic Himalayan rivers of the Indo-Gangetic alluvial plain, the Godavari and its tributaries are incised in rock or alluvium and have relatively stable sandy channels. As for other rivers affected by storms (Edwards and Owens, 1991; Hilton et al., 2008), extreme rainfall events are disproportionately important for erosion in the Godavari watershed and in subsequent transport of sediments to the ocean (Kale, 2003). Given their incised morphology, shifts in channel position in response to floods are, however, rare above the Godavari Delta (Kale, 2002). Floodplains are limited in extent (2 % of the basin; Bikshamaiah and Subramanian, 1988), and loss of sediments to overbank deposition is minor (Kale, 2002). Therefore storage is minimal in these intermediate alluvial reservoirs that normally would increase the residence time of sediments, including particulate organic carbon.

Once reaching the Bay of Bengal, sediment delivered by the Godavari has constructed a large Holocene delta (Rao et al., 2005; Cui et al., 2017). Offshore from the Godavari mouth, a persistent sediment plume extends over 300 km during the monsoon season, when over 90 % of the fluvial sediment is discharged (Sridhar et al., 2008). Because the shelf in front of the delta is unusually narrow (i.e., under 10 km at our core location) copious sediment deposition takes place directly on the continental slope, resulting in sediment accumulation rates as high as 250 cm kyr^{-1} (Ponton et al., 2012). Owing to the narrow shelf, changes in sea level would also have minimal effects on sediment deposition at our site, especially after the early Holocene when the global sea level reached within a few meters of modern values (Lambeck et al., 2014). For these reasons our core located close to the river mouth ($\sim$ 35 km) is unlikely to contain any significant contributions from other sediment sources, in agreement with previous studies (e.g., Bejugam and Nayak, 2017).

The relatively simple sedimentary regime of the Godavari system in combination with the monsoon-dominated climatology and simple geology of the Godavari Basin allows for relatively straightforward interpretation of sediment sources and transfer processes. The monsoon wash load is rapidly and directly delivered to the continental margin without significant trapping in intermediate depocenters along the river. As the suspended load makes up over 95 % of the total sediment transported by the Godavari (Syvitski and Saito, 2007), the wash-load-derived terrestrial proxies are representative of the production of fine-grained sediment in the basin. Potential contributions from resuspension of shelf sediments cannot be excluded but are likely minor due to the narrowness of the shelf; furthermore, given the large sedimentation rates on the shelf itself (Forsberg et al., 2007), the resuspended sediment is expected to be quasi-contemporaneous with sediments arriving on the slope directly from the river plume.

3 Hydroclimate in the Core Monsoon Zone

We have previously reconstructed the Holocene paleoclimate using the same sediment core discussed herein (Ponton et al., 2012; Zorzi et al., 2015). Terrestrial reconstructions were based on the carbon isotopic compositions of higher plant leaf-wax biomarkers (i.e., long-chain *n*-alkanoic acids C_{26-32}) and pollen, whereas contemporaneous sea surface paleoceanographic conditions in front of the Godavari Delta came from the oxygen isotopic composition of planktonic foraminifer *Globigerinoides ruber*. Sedimentary leaf waxes provide an integrated δ^{13}C record of the flora in the CMZ that document an increase in aridity-adapted vegetation (C_4 plants) after the monsoon maximum in the early Holocene (Ponton et al., 2012; Fig. 2). The overall trend of the δ^{13}C leaf-wax record agrees with the view that the seasonality of Northern Hemisphere insolation (Ponton et al., 2012) led to progressively weaker monsoons over the Holocene. However, two clear aridification steps are evident: between $\sim$ 5000 and 4500 years ago, and $\sim$ 1700 years ago (Fig. 2). Pollen from the same core (Zorzi et al., 2015) reinforces these conclusions: coastal forest and mangrove pollen (Fig. 2) that are typical for the more humid coastal regions

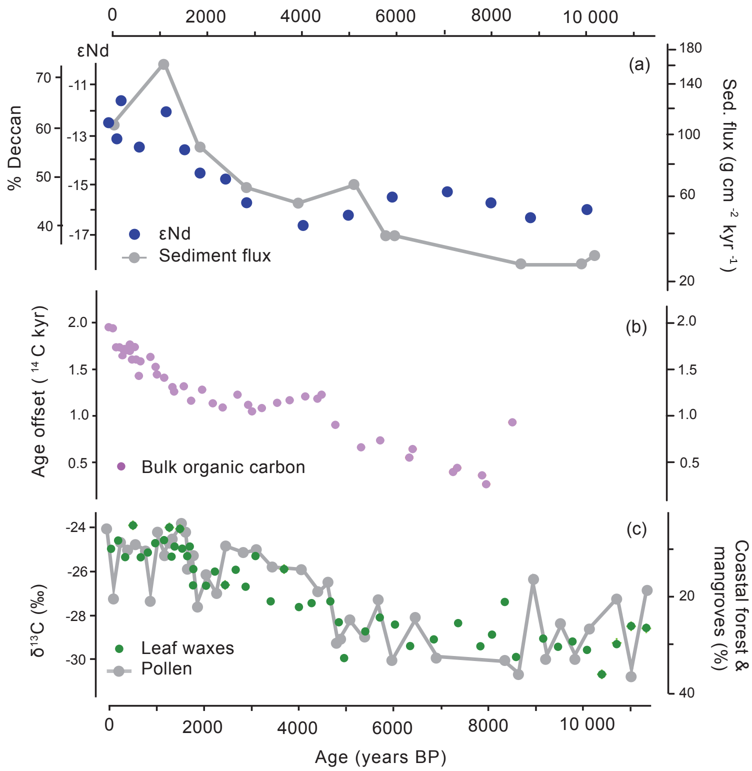

Figure 2. Paleoenvironmental reconstructions from core NGHP-01-16A for the Core Monsoon Zone as integrated by the Godavari River: **(a)** sediment fluxes as mass accumulation rates and sediment sources from Nd isotope fingerprinting (Deccan Trap sediment contribution is estimated from a two-end-member model; see text and Supplement); **(b)** TOC radiocarbon age offset relative to depositional age; **(c)** hydroclimate and ecology from pollen (Zorzi et al., 2015) and leaf-wax carbon isotopes (Ponton et al., 2012).

of the Eastern Ghats and Godavari Delta declined over the Holocene.

Dryness-adapted thornbush pollen from the Deccan Plateau increased substantially after the second aridification step ∼ 1700 years ago, overlapping well with the maximum contribution of C_4 plant-derived leaf waxes (see Zorzi et al., 2015). For the same time interval, the ice-volume-corrected oxygen isotopic composition of planktonic foraminifer *Globigerinoides ruber* documented a series of low values interpreted as high-salinity events at the Godavari mouths (see Ponton et al., 2012). Together these continental and oceanic records suggest that the CMZ aridification intensified in the latest Holocene via a series of short drier episodes (Ponton et al., 2012). This interpretation is reinforced by speleothem-derived records from central and northern India for the past thousand years (Sinha et al., 2011), and the overall evolution of the CMZ hydroclimate as seen from our core is supported by local reconstructions from the Lonar crater lake in central India (Prasad et al., 2014; Sarkar et al., 2015), Godavari Delta (Cui et al., 2017) and other records from the larger Indian monsoon domain (Gupta et al., 2003; Fleitmann et al., 2003; Prasad and Enzel, 2006; Berkelhammer et al., 2012; Dixit et al., 2014).

4 Erosion in the Godavari Basin

The Holocene sediment flux at our core location (Fig. 2) is representative of the Godavari continental slope (Mazumdar et al., 2009; Ramprasad et al., 2011; Joshi et al., 2014) and is driven by changes in the siliciclastic sedimentation rate as dilution by biogenic carbonates is less than 5 % (Johnson et al., 2014). Despite a lower sea level at the time, the flux was relatively small in the early Holocene (∼ 25 g cm^{-2} kyr^{-1}) but began to increase after 6000 years ago (∼ 40 g cm^{-2} kyr^{-1}), as soon as the monsoon started to decline but well before the adoption of Neolithic agriculture and settlement of the savannah zone of the central peninsula (∼ 4500 years ago; Fuller, 2011). Between 4000 and 3500 years ago permanent agricultural settlements spread throughout the Deccan Plateau. The associated small-scale metallurgy (copper working) requiring firewood together with the agricultural intensification probably also affected erosion via widespread deployment of two cropping seasons (Kajale, 1988; Fuller and Madella, 2001). As the climate remained arid, sediment fluxes stayed high despite a phase of agricultural abandonment and depopulation between ∼ 3200 and 2900 years ago (Dhavalikar, 1984; Roberts et al., 2016).

A further step increase in the sediment flux ($\sim 90\,\mathrm{g\,cm^{-2}\,kyr^{-1}}$ on average) occurred after ~ 3000 years ago, this time with no apparent concurrent change in climate. The Nd isotopic signal points to an increase in the Deccan sedimentary output at the time, after a muted variability earlier in the Holocene when the Indian Craton consistently provided ~ 50–60 % of the sediments (Fig. 2; see Supplement). Ferrimagnetic minerals interpreted as originating from the Deccan Plateau (Sangode et al., 2001; Kulkarni et al., 2014) also increase in late Holocene sediments in the Godavari Delta (Cui et al., 2017) and Bay of Bengal (Kulkarni et al., 2015), supporting our interpretation. Augmented Deccan input was suggested for the Godavari Delta even earlier after ~ 6000 years ago (Cui et al., 2017), in step with the initial aridification.

New improvements in agricultural technology became widespread in the Deccan Plateau, including use of iron agricultural tools (Mohanty and Selvakumar, 2001) that required firewood-fueled smelting (Fuller, 2008). A new phase of agricultural settlement began in the middle Godavari Basin (eastern Maharashtra) between ~ 3000 and ~ 2800 years ago (Brubaker, 2000). However, the largest boost in sediment flux occurred after ~ 2000 years ago, when the monsoon reached its driest phase and when further increases in population occurred, resulting in the founding of towns and the first cities of the region at the beginning of the Indian Historic Period (Allchin, 1995; Parabrahma Sastry, 2003). This doubling in sediment flux relative to the early Holocene values involved a basin-wide increase in erosion. The contribution from the Deccan Plateau, although at its maximum according to the Nd isotope mixing model, only accounts for a 15 % shift in sediment source (Fig. 2).

Overall, watersheds with high precipitation have higher discharge, and discharge magnitude is considered a primary regulator for sediment and carbon erosional fluxes to the ocean (e.g., Summerfield and Hulton, 1994; Ludwig et al., 1996; Galy et al., 2015). However, our Godavari record shows that erosional output is maximized by aridity because significant rain and seasonal floods still occur during the summer monsoon season (Mujumdar et al., 1970; Kale, 2003). Aridification and/or agricultural expansion lead to changes in vegetation type (i.e., forest decrease in favor of savannah) and cover (i.e., shrinking of naturally vegetated lands in favor of agricultural and/or degrading arid lands) that exacerbate soil erosion (i.e., Langbein and Schum, 1958; Dunne, 1979; Walling and Webb, 1983; Istanbulluoglu and Bras, 2005; Vanacker et al., 2007; Collins and Bras, 2008).

5 Carbon export from the Godavari Basin

The terrigenous organic carbon exported by rivers consists of a mixture of dissolved and particulate components derived from contemporary vegetation and of carbon stored in bedrock, soils and fluvial sediments that may be significantly pre-aged (Smittenberg et al., 2006; Galy and Eglinton, 2011; Feng et al., 2013). On the Godavari slope, the terrigenous fraction dominates the total organic carbon (TOC) in marine sediments (Johnson et al., 2014). In agreement with this, TOC radiocarbon ages in our core have been previously found to be remarkably similar to co-located ages of the strictly terrigenous higher plant leaf-wax fraction (Ponton, 2012). This age similarity also excludes interferences from within-river biological productivity (e.g., Eglinton and Hamilton, 1967; Eglinton and Eglinton, 2008). To assess the variability of the terrigenous carbon age exported by Godavari River based on this understanding we used high-resolution TOC radiocarbon measurements to calculate radiocarbon age offsets relative to the atmosphere (Soulet et al., 2016; see Supplement). Over the Holocene, these biospheric organic carbon radiocarbon age offsets in our core mirror the history of erosion in the basin (Fig. 2).

As a first-order observation, TOC ages (Fig. 2 and Supplement) are significantly older (~ 200 to 2000 ^{14}C years) than their depositional age in our Godavari core. Before 5000 years ago the bulk organic carbon radiocarbon age offset was ~ 600 ^{14}C years old on average. In contrast, the highly erosional regime under both climatic and early human pressure in the late Holocene led to the export of significantly older carbon from the terrestrial biosphere, i.e., ~ 1300 ^{14}C on average. This increase in radiocarbon age offset occurred largely during the two aridification steps identified by Ponton et al. (2012): more abruptly between ~ 5000 and 4500 years ago and more gradually after ~ 1700 years ago (Fig. 2).

In the absence of significant storage in alluvial sediments in the Godavari catchment, several processes can explain the doubling in age offset over the Holocene: an overall slowing of soil carbon turnover in the drying climate of central India, a decrease in TOC contribution from contemporaneous vegetation relative to older (pre-aged) soil carbon input and/or deeper exhumation of soils contributing increasingly older carbon. Given the drastic changes in vegetation cover and increase in erosion in the Godavari Basin, a decrease in soil turnover is unlikely during the Holocene aridification process (Carvalhais et al., 2014). In turn, the good agreement between the pollen and leaf-wax δ^{13}C records in our core (Ponton et al., 2012; Zorzi et al., 2015) with independent monsoon reconstructions suggests sustained delivery of recently fixed biospheric organic carbon to the delta. Thus, the doubling in age offset over the Holocene is best explained by increasing contributions from an older soil component, which could only come through deeper erosion. Because the age of soil organic carbon in soil profiles increases with depth (Trumbore, 2009), older mixtures imply a deeper soil erosion, whether uniform or through gullies, which are common especially on the Deccan Plateau (Kothyari, 1996).

6 The monsoon erosional pump

Overall, these multiple lines of evidence indicate that soil erosion in the CMZ, as integrated by the Godavari River, increased throughout the basin immediately as climate began to dry at the end of the mid-Holocene and was further enhanced by Deccan agricultural activities in the late Holocene. The likely mechanism for this erosion acceleration is the extreme seasonal distribution of the rainfall that characterizes the monsoon (Wang and Ding, 2008), which promoted erosion on the more sparsely vegetated landscapes (Molnar, 2001; DiBiase and Whipple, 2011; Plink-Björklund, 2015). Our findings thus point to a veritable "monsoon erosional pump" that accelerates during minimum land cover conditions when the protective role of vegetation is reduced, whether naturally or by humans. The volume of total eroded sediments since the mid-Holocene must have been considerable as the continental margin growth accelerated with the shelf edge aggrading ~ 80 m in the last ~ 2000 years alone (Forsberg et al., 2007).

This "land cover mode" of the monsoon erosional pump must have been active before the Holocene as well, affecting the transfer of terrigenous sediment, solutes and carbon from land to the ocean. The beat of monsoon precipitation on orbital timescales is not well constrained but considered to be modulated by at a combination of precession and obliquity frequencies based on monsoon wind reconstructions (e.g., Clemens and Prell, 2003). Such complex variability did not inevitably follow the sea level cyclicity (e.g., Goodbred and Kuehl, 2000), which is usually assumed to control most of the sediment transfer from land to the deep ocean (see Blum and Hattier-Womack, 2009, and references therein for an analysis underlining the increased recognition for a climate role). Thus, untangling the effects of the monsoon is difficult, especially during the Quaternary (e.g., Phillips et al., 2014), but may be easier to discern earlier when the sea level change magnitude was reduced. Land cover effects are less likely to occur in the upper basins of Himalayan monsoonal rivers where there are other sources of water such as snow or glaciers and where elevation (i.e., temperature) and orographic precipitation promote ecological stability (Galy et al., 2008a). The erosional pump in these high, steep regions is still active due to monsoonal seasonality but in a "topographic mode" dominated primarily by landslides (Montgomery and Brandon, 2002; but see Olen et al., 2016, for an alternative viewpoint). However, the land cover mode for the erosional pump should still be active in their lower basins where aridity controls vegetation type and cover (e.g., Galy et al., 2008b).

Recent coupled erosion–carbon cycling modeling suggests that long-term anthropogenic acceleration of erosion has had a significant impact on the global carbon cycle by intensifying the burial of terrigenous carbon (Wang et al., 2017). Prior to damming, the monsoon domain supplied $\sim 70\,\%$ of the sediment load coming from large rivers (Syvitski and Saito, 2007), although it only covers $\sim 15\,\%$ of the Earth's surface (Hsu et al., 2011). Therefore, we suspect that the cumulative effect of the monsoon erosional pump on the carbon budget was substantial in augmenting the burial of terrigenous carbon during the Holocene and needs to be estimated for inclusion in assessments of the net soil–atmosphere carbon exchange.

Data availability. The Supplement contains all data discussed in this paper, including radiocarbon and Nd isotope measurements.

The Supplement related to this article is available online at https://doi.org/10.5194/esurf-5-781-2017-supplement.

Competing interests. The authors declare that they have no conflict of interest.

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Short communication: Massive erosion in monsoonal central India linked to late Holocene land cover degradation

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Supplementary Materials

The detrital fraction provenance was assessed using Nd isotopic ratios (Supplementary Table 1). Nd chemistry was done with conventional ion chromatography following the method of Bayon et al. (2002). Nd analyses were performed on the NEPTUNE multi-collector ICP-MS at WHOI with the internal precision of 5-10 ppm (2 sigma). The external precision, after correction to value for LaJolla standard ($^{143}Nd/^{144}Nd$=511847) is approximately 15 ppm (2 sigma). $^{143}Nd/^{144}Nd$ isotopic composition is expressed here as εNd (DePaolo and Wasserburg, 1976) units relative to ($^{143}Nd/^{144}Nd$)CHUR= 0.512638 (Hamilton et al., 1983). Very low εNd values are generally found in continental crusts, whereas higher (more positive) εNd values are commonly found in mantle-derived melts (DePaolo, 1988), such as those of large igneous provinces.

The average εNd for the Deccan basalts is +1 ± 5 and for the Indian craton is -35 ± 8 (GEOROC Database, Geochemistry of Rocks of the Oceans and Continents, Max Plank Institute for Chemistry, Mainz, Germany, http://georoc.mpch-mainz.gwdg.de/). The measured εNd value of a sample was expressed as a simple mixture of sediment derived from the two end-members:

$$\varepsilon \mathrm{Nd}_{\,Sample} = f \cdot \varepsilon \mathrm{Nd}_{\,Deccan} + (1-f) \cdot \varepsilon \mathrm{Nd}_{\,Craton}$$

Where (f) is the fraction of Deccan derived sediments, $(1-f)$ the fraction of Craton derived sediments in the mixture, and f is a number between 0 and 1.

Sediment fluxes (Supplementary Table 2) were constructed as mass accumulation rates assuming negligible carbonate inputs (Johnson et al., 2014) using measured dry bulk densities on the samples used for foram radiocarbon dating and sedimentation rates from the age model of Ponton et al. (2012).

The high resolution series of bulk TOC ^{14}C content was measured at the Geological Institute and the Laboratory of Ion Beam Physics, ETH Zürich (Supplementary Table 3). The bulk TOC ^{14}C measurements made at ETHZ are detailed in McIntyre et al. (2016). Duplicates of 70-90 mg of freeze-dried sediment samples were weighed in pre-combusted silver boats (Elementar) and fumigated with HCl to remove carbonate (Komada et al., 2008). The samples were subsequently neutralized and dried over solid NaOH pellets to remove residual acid. The samples were then wrapped in a second tinfoil boats (Elementar9 and pressed prior to analysis.

Samples were graphitized by the automated graphitization equipment (AGE) and analysed for ^{14}C using the MICADAS system (Ionplus) and an ampoule cracker system following the procedure outlined in Wacker et al. (2013). The other batch was then run as gas on the coupled EA-IRMS-AMS system at ETHZ. The data for the TOC ^{14}C content showed that samples analysed using graphite and CO_2 are within 2σ of each other (McIntyre et al., 2016).

For microscale (≤ 20 μg C) AMS ^{14}C analysis, comprehensive procedural blank assessment is critical in order to constrain analytical uncertainty (Drenzek, 2007; Santos et al., 2010; Tao et al., 2015). An evaluation of the complete procedure used here (chemical extraction, derivatization, PCGC isolation, final clean-up and combustion steps) yielded a procedural blank of 1.2±0.4 μg C per 30 PCGC injections, with an $\Delta^{14}C$ of –382±126‰ (Tao et al., 2015). Separate assessment of modern and fossil C blanks yielded 0.8±0.2 μg of modern C contamination (i.e., $\Delta^{14}C$ = 0‰) and 0.5 ± 0.1 μg of dead C contamination (i.e., $\Delta^{14}C$ = –1000‰), with a combined procedural blank of 1.3±0.2 μg C per PCGC 30 injections with a $\Delta^{14}C$ value of –325±129‰. From this assessment as well as a previous assessment (Drenzek, 2007), we estimate that the analytical uncertainty for ^{14}C analysis of FAs ranges from 6 to 40‰ (ave., 12‰).

The raw and calibrated radiocarbon age models used to estimate depositional ages are from Ponton et al. (2012). The age of the bulk TOC at the time of their deposition was estimated by taking the offsets between their radiocarbon content and the interpolated reservoir-corrected foraminifera-based radiocarbon age (Supplementary Table 3). The reservoir correction used was 400 years. Taking a conservative approach we calculated the propagated error for the radiocarbon age offsets (Supplementary Table 3) as:

$$err.\ offset = ((err.\ TOC\ ^{14}C\ measurement)^2 + (max.\ err.\ foram\ ^{14}C\ measurement)^2)^{1/2}$$

where the maximum error for the foraminifera ^{14}C measurements used in the age model was 55 ^{14}C years (Ponton et al., 2012). The resulting errors for the offset range between 63 and 80 years.

Supplementary Table 1. Downcore measurements of $^{143}Nd/^{144}Nd$ composition with corresponding εNd for the Holocene section of NGHP-01-16A in front of Godavari delta.

Depth (mbsf)	Age (yr)	143Nd/144Nd	εNd
0.00	59	0.511999	-12.46
0.16	159	0.511973	-12.97
0.32	256	0.512035	-11.76
0.80	542	0.511946	-13.50
1.70	1085	0.512018	-12.09
2.50	1627	0.511945	-13.52
3.00	2019	0.511888	-14.63
3.60	2567	0.511888	-14.63
4.00	2990	0.511830	-15.76
4.80	4002	0.511780	-16.74
5.40	4936	0.511809	-16.17
6.00	6043	0.511847	-15.43
6.50	7116	0.511856	-15.25
6.90	8082	0.511832	-15.72
7.20	8873	0.511801	-16.33
7.60	10024	0.511822	-15.92

Supplementary Table 2. Downcore estimates of sediment fluxes for the Holocene section of NGHP-01-16A in front of Godavari delta based on calibrated foraminifera ^{14}C depositional ages.

Depth (mbsf)	**Age** (yr)	**Sediment Flux** ($g/cm^2/kyr$)
0.075	0	87.0
1.475	1104	151.2
2.975	1852	70.7
4.045	2895	47.8
4.775	4046	41.8
5.355	5331	49.2
6.015	5996	31.0
6.435	6198	31.0
7.215	9056	23.7
7.655	10314	23.7
7.885	10619	25.8

Supplementary Table 3. Downcore measurements of bulk TOC measured at ETH and offsets to foram ^{14}C depositional ages for the Holocene section o f NGHP-01-16A.

Depth (mbsf)	Age (kyr)	^{14}C Depositional Age (kyr)	TOC ^{14}C Age (kyr)	Error TOC ^{14}C Age (kyr)	^{14}C Age Offset (kyr)	Max. Error ^{14}C Age Offset (kyr)
0.265	223	31	1844	51	1813	75
0.350	274	117	1928	32	1811	63
0.440	328	205	1833	32	1627	63
0.485	355	249	1869	32	1620	63
0.545	391	305	1924	51	1619	75
0.660	459	410	1948	51	1538	75
0.755	516	494	2098	51	1604	75
0.840	566	567	2206	51	1639	75
0.890	595	609	2251	32	1642	63
0.940	625	650	2238	32	1588	64
1.035	681	726	2246	51	1519	75
1.085	711	766	2389	51	1623	75
1.145	746	812	2306	32	1494	64
1.235	800	881	2233	32	1352	64
1.305	842	933	2421	51	1488	75
1.635	1044	1166	2709	33	1542	64
1.825	1164	1293	2712	52	1419	75
1.865	1190	1320	2673	32	1354	64
2.060	1318	1446	2783	32	1337	64
2.315	1493	1607	2850	52	1243	76
2.365	1529	1638	2837	32	1198	64
2.640	1732	1811	3069	52	1258	76
2.835	1884	1935	3033	52	1098	76
3.090	2096	2100	3327	52	1227	76
3.315	2295	2253	3353	32	1101	64
3.365	2341	2287	3364	52	1077	76
3.560	2528	2427	3466	32	1039	64
3.855	2831	2652	3835	33	1183	64
3.915	2896	2700	3842	53	1142	76
4.085	3086	2842	3902	53	1060	76
4.130	3138	2881	3882	52	1001	76
4.310	3354	3042	4086	33	1044	64
4.575	3693	3299	4386	53	1088	76
4.745	3925	3476	4590	54	1114	77
4.965	4243	3723	4871	34	1148	65
5.140	4511	3934	5063	34	1129	65
5.180	4574	3984	5141	33	1157	64
5.355	4860	4213	5085	54	872	77
5.655	5384	4640	5301	34	661	65
5.865	5778	4969	5702	34	733	65
6.165	6382	5482	6063	34	581	65
6.200	6455	5546	6210	55	664	78
6.575	7290	6278	6704	56	427	78
6.605	7360	6340	6810	35	470	65
6.825	7893	6820	7218	56	399	79
6.860	7981	6899	7208	35	308	65
7.065	8510	7384	8276	58	892	80

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