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Pliocene expansion of C_4 vegetation in the core monsoon zone on the Indian Peninsula

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Abstract. The expansion of C_4 vegetation during the Neogene was one of the largest reorganizations of Earth's terrestrial biome. Once thought to be globally synchronous in the late Miocene, site-specific studies have revealed differences in the timing of the expansion and suggest that local conditions play a substantial role. Here, we examine the expansion of C_4 vegetation on the Indian Peninsula since the late Miocene by constructing a ~6 million year paleorecord with marine sediment from the Bay of Bengal at Site U1445 drilled during International Ocean Discovery Program Expedition 353. Analyses of element concentrations indicate the marine sediment originates from the Mahanadi River in the Core Monsoon Zone (CMZ) of the Indian Peninsula. Hydrogen isotopes of the fatty acids of leaf waxes reveal an overall decrease in the CMZ precipitation since the late Miocene. Carbon isotopes of the leaf wax fatty acids suggest C_4 vegetation on the Indian Peninsula existed before the end of the Miocene, but expanded to even higher abundances during the mid-Pliocene to mid-Pleistocene (3.5 to 1.5 Ma). Similar to the CMZ on the Indian Peninsula, a Pliocene expansion or re-expansion has previously been observed in northwest Australia and in East Africa, suggesting that these tropical ecosystems surrounding the Indian Ocean remained highly sensitive to changes in climate after the initial spread of C_4 plants in late Miocene.

1. Introduction

The expansion of plants using the C_4 photosynthetic pathway is one of the most dramatic reorganizations of the global biome during the Neogene. A widespread late-Miocene expansion (8 to 6 Ma) is well documented and many studies have interpreted the broadly synchronous timing as ecosystems adapting to decreasing pCO_2 (e.g., Ehleringer et al., 1991; Ehleringer and Cerling, 1995; Cerling et al., 1993, 1997; Herbert et al., 2016). However, an increasing number of studies have shown that the timing, regional patterns, rate and drivers of C_4 grassland expansion were much more diverse and complex. Along with low pCO_2, a C_4 photosynthetic pathway is better adapted to higher temperature, aridity, seasonality, and during disturbances such as flood, droughts, and fires (e.g., Edwards et al., 2010, and references therein). The interplay

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of these parameters varies amongst regions. Resolving the precise timing and factors leading to major changes in vegetation demands site-specific studies (Strömberg et al., 2011; Zhou et al., 2014).

Our study provides a novel piece of the puzzle in unraveling the complexities of C_4 expansion by constructing a 6 million year (Myr) record of C_4 vegetation and aridity on the Indian Peninsula. The marine sediment record is from International Ocean Discovery Program (IODP) Site U1445 (17°44.72'N, 84°47.25'E; 2,503m water depth; Fig. 1A) drilled in the Bay of Bengal (BoB) close to the mouths of the Mahanadi River. Lithologies at Site U1445 include calcareous fossils, biosilica, silt, and clays (including glauconite), and are overall described as hemipelagic sediment (Clemens et al., 2016). The Indian Monsoon dictates climate patterns in the Mahanadi River drainage basin: rainy summers, dry winters, and an annual reversal of wind direction (Gadgil, 2003; Sarkar et al., 2015). Highly sensitive to the seasonal changes, more than 80% of runoff from the Mahanadi River into the BoB occurs during the summer (Chakrapani and Subramanian, 1990).

Previous reconstructions of Neogene C_4 expansion in regions affected by the Indian Monsoon use deposits originating in the Himalayas and their piedmont regions (France-Lanord and Derry, 1994; Quade and Cerling, 1995; Quade et al., 1995; Cerling et al., 1997; Freeman and Colarusso, 2001; Sanyal et al., 2004; Behrensmeyer et al., 2007; Galy et al., 2010; Ghosh et al., 2017). The Mahanadi River drains a relatively low-elevation region of the Indian Peninsula distinct from the nearby mountain ranges (e.g., the Western Ghats, the Himalaya, Indo-Burman ranges Fig. 1, Xie et al., 2006). With minimal orographic precipitation in the Mahanadi River basin, rainfall in this "Core Monsoon Zone" (CMZ) represents the mean behavior of the Indian Monsoon (Fig. 1; Ponton et al., 2012; Sarkar et al., 2015; Giosan et al., 2017,and references therein).

Although agriculture dominates present-day vegetation, models predict natural flora of the Mahanadi basin would be closed-canopy, moist deciduous forests and moist-to-dry woodlands with rare open spaces (Fig. 1C, Zorzi et al., 2015 and references therein). Today the region encompasses a range of C_3 and C_4 vegetation, but proxies and models suggest that the plant communities are highly sensitive to glacial-interglacial changes with nearly all flora utilizing a C_4 pathway during the last glacial maximum (Galy et al., 2008; Phillips et al., 2014; Zorzi et al., 2015, and references therein). The behavior of vegetation in the CMZ over million-year timescales is unknown.

Here, we use inorganic bulk geochemical analyses to fingerprint the origin of sediment at Site U1445 to be from the Mahanadi River. Then we use bulk organic and compound-specific biomarkers at the same site, including carbon and hydrogen isotope measurements of leaf wax fatty acids, to reconstruct the changes in C_4 vegetation and rainfall in the CMZ of the Indian Peninsula over the last ~6 Myr (Fig. 2).

2. Methods

We constructed an age model for Site U1445 using CLAM software in R (Blaauw, 2010) to fit a locally weighted spline to biostratigraphic and magnetostratigraphic ages (Clemens et al., 2016; Fig. S1). Our samples were collected from the same hole in which the age model was constructed. We measured major, trace, and rare earth element concentrations on

30 bulk sediment samples spanning 0-6 Ma to determine sediment provenance (Dunlea et al., 2015; Appendix A). Then we analyzed bulk organics (Appendix B) as well as compound-specific biomarkers (Appendix C) from 57 samples to reconstruct hydrological and vegetation changes. Samples for biomarker analysis were collected from Site U1445 in pairs, visually targeting relatively light and dark layers at similar depths to capture the variability range on shorter timescales while characterizing longer trends.

3. Results

The age model constructed at Hole U1445A suggests that long-term sedimentation rates have been overall continuous and fairly constant (Fig. S1). Shipboard scientists observed thin turbiditic sequences (~2-20cm thick) throughout Site U1445 and the expansion and dissociation of gas hydrates upon recovery that may muddle a higher-resolution record (Clemens et al., 2016). However, Site U1445 has fewer and smaller turbidite deposits relative to other sites drilled in this region and the records spanning million-year timescales are likely relatively undisturbed.

To determine the provenance of the aluminosilicate fraction, we examined the proportions of Al, Ti, Sc, Nb, La, and Th concentrations, because other elements (e.g., Fe, K, Mg, Si, Zr, Hf) may be affected by continental weathering, sorting during transport, and post-depositional authigenic processes. The results from 30 samples have almost constant element proportions of the selected elements, indicating that the aluminosilicate fraction of sediment not significantly varied over the past 6 Myr. The composition of the 30 samples, even amongst the light and dark layers, matches the composition of lithologies that comprise the Mahanadi basin such as Precambrian granite and gneisses of the Indian craton and associated sedimentary deposits (Sharma, 2009; Fig. S2; Table S1). Marine sediment deposits in other parts of the Bay of Bengal closer to the Krishna and Godavari Rivers or Ganges-Brahmaputra Rivers have a more mafic or highly variable composition that is not observed at Site U1445 (Tripathy et al., 2014; Fig. S2). As such, we interpret our results as recording terrestrial changes in the CMZ, specifically the Mahanadi drainage basin.

The pairs of samples used for organic analyses are spaced ~28 m apart (~260 kyr intervals) and the adjacent light and dark layers within each pair were 0.2 m to 4.3 m apart in the sediment core (2 kyr to 46 kyr; Fig. S1). The color difference can be related to the total organic carbon content (wt%; TOC) content with darker layers having 1.0 to 2.8 times more than adjacent lighter layers (Fig. 2A; Fig. S3; Table S2).

Long-chain normal fatty acids of leaf waxes are derived from land plants and are well preserved during transport and burial in marine sediment (Eglinton and Eglinton, 2008). We focus on the C_{30} chain length to avoid possible contaminations from non-terrestrial sources that contribute shorter chain length fatty acids (Fig. S4; Table S3). The results of the $\delta^{13}C$ of C_{30} fatty acid of leaf waxes ($\delta^{13}C_{FA}$) show a 5‰ increase from mid-Pliocene to mid-Pleistocene (3.5 to 1.5 Ma), after which $\delta^{13}C_{FA}$ decreases and becomes more variable from 1.5 Ma to the present (Fig. 2B). The hydrogen isotope compositions of the leaf wax fatty acids (δD_{FA}) increase over the past 6 Myr, but have a wide range amongst light and dark

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layers and shorter time intervals (Fig. 2C; Table S3). Before the mid-Pliocene (3.5 Myr) δD_{FA} ranges from -177‰ to -146‰ and after the mid-Pleistocene (1.5 Myr) the δD_{FA} increases to between -163‰ to -125‰ (Fig. 2C).

4. Discussion

4.1 C_4 Expansion on the Indian Peninsula

The $\delta^{13}C_{FA}$ of terrestrial plants is primarily a function of the photosynthetic pathway and isotopic composition of atmospheric CO_2 (e.g., Farquhar et al., 1989). In this study, the 5‰ increase in $\delta^{13}C_{FA}$ is greater than the reconstructed $\delta^{13}C$ of atmospheric CO_2 (≤1‰; Tipple et al., 2010), suggesting that a correction for $\delta^{13}C_{CO2}$ would only slightly adjust our results. Thus we interpret $\delta^{13}C_{FA}$ as reflecting the amount of C_4 relative to C_3 vegetation produced in the CMZ.

Approximating typical $\delta^{13}C_{FA}$ of C_3 and C_4 plants (Chikaraishi et al., 2004; Ponton et al., 2012), we estimate that 36% to 68% (avg. 56% ± 9% s.d.) of the vegetation in the CMZ utilized a C_4 photosynthetic pathway from 6 Ma until 3.5 Ma (Fig. S5). Thus, the environmental threshold for C_4 photosynthetic pathway had already been crossed before the end of the late Miocene. Later in the mid-Pliocene, the reconstruction shows another distinct expansion reaching 62% to 88% (avg. 80% ± 8% s.d.) of C_4 vegetation in the early Pleistocene (Fig. 2B). The substantial change in vegetation from 3.5 to 1.5 Ma suggests multiple steps of C_4 expansion in the CMZ, rather than a singular late-Miocene expansion. From 1.5 Ma to the present, the average proportion of C_4 vegetation decreased and became more variable (54% to 87%, avg. 71% ± 11% s.d.; Fig. 2B), which may reflect the sensitivity of the region to glacial-interglacial variations observed in shorter records from this region (e.g., Zorzi et al., 2015, and references therein).

4.2 Aridification of the Indian Peninsula

The amount of precipitation, mixing of different air masses, and plant physiology can each vary the hydrogen isotopic composition of leaf wax fatty acids (δD_{FA}; e.g., Eglinton and Eglinton, 2008). The mixing of two air masses with unique δD values was recently observed to drive δD of rainfall in New Delhi, India, but, similar to the amount of precipitation, the relatively depleted δD corresponded with wetter conditions (Hein et al., 2017). Thus, we corrected for the physiological effects of C_3 versus C_4 plants on δD_{FA} (Fig. S5; Fig. 2C), and interpreted the corrected δD_{FA} as a qualitative proxy for aridity or the relative amount of precipitation. The δD_{FA} results suggest an overall drying of the CMZ on the Indian Peninsula over the past 6 Myr. The shorter-term scatter in the δD_{FA} record may reflect higher frequency variations in aridity or rainfall.

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4.3 Global Patterns of C_4 Expansion

Untangling the triggers of C_4 expansions during the Neogene requires compiling records from many sites to identify global versus local trends. Here, we compare our record with other compound-specific biomarker records of C_4 expansion at sites in the Indian Ocean or adjacent land and seas.

Multiple proxy records document a late-Miocene C_4 expansion in the Ganges or Brahmaputra River basins such as the Siwalik Group in northern Pakistan or BoB sites receiving outflow sediment (Fig. 3A France-Lanord and Derry, 1994; Quade and Cerling, 1995; Cerling et al., 1997; Freeman and Colarusso, 2001; Sanyal et al., 2004; Behrensmeyer et al., 2007; Ghosh et al., 2017). Collectively, the reported timing of C_4 expansions in the Himalaya region ranges from 9 to 5 Ma, most commonly 8 to 6 Ma (Behrensmeyer et al., 2007). Rather than a uniform timing, detailed sampling of various deposits around the Siwalik regions shows that C_4 vegetation expansion was staggered amongst nearby sub-environments with different local conditions (Ghosh et al., 2017, and references therein). Another biomarker record documents a late-Miocene expansion in a wide continental region north and west of the Arabian Sea (Site 722; Fig. 1A, Fig. 3B; Huang et al., 2007). Once C_4 vegetation expanded at each of these sites, the records suggest there is overall little change in the amount of C_4 vegetation after the late Miocene.

In contrast, the CMZ of the Indian Peninsula and a few other records around the Indian Ocean document an expansion of C_4 vegetation during the Pliocene (Fig. 3). Marine deposits in the Gulf of Aden originate from northeast Africa and record a late-Miocene C_4 expansion, followed by a relapse to predominantly C_3 vegetation ~4.3 Ma and a re-expansion of C_4 plants in the Pliocene (Site 231; Fig. 1A, Fig. 3C; Feakins et al., 2005, 2013; Liddy et al., 2016). The Pliocene re-expansion is consistent with other records from tropical East Africa (e.g., Levin et al., 2004; Cerling et al., 2011). A C_4 expansion in the Pliocene is also observed in northwest Australia (Site 763A; Fig. 3E; Andrae et al., 2018). There is little evidence of significant C_4 vegetation prior to the Pliocene, suggesting a relatively late onset of C_4 vegetation expansion in northwest Australia (Fig. 3).

Collectively, a significant regional expansion in the Pliocene, distinctly after the first late-Miocene expansion, is common at least amongst tropical East Africa, Northwest Australia, and the Indian Peninsula. Farther from the Indian Ocean, there is also evidence that East and Central Asia also experienced multiple steps of C_4 expansion through the Pliocene, depending on local sub-climates (e.g., An et al., 2005; Passey et al., 2009; Zhou et al., 2014). The modes of climate variability around the Indian Ocean likely differed throughout the Pliocene and may have set the stage for a regional, multi-step reorganization of the terrestrial biome.

4.4 Triggers of C_4 Expansion in the Pliocene

Many studies hypothesize that the widespread expansion of C_4 vegetation in the late Miocene was triggered by pCO_2 decreasing below a temperature-dependent threshold (Herbert et al., 2016 and references therein). Since C_4-dominated ecosystems are favored under low pCO_2 and higher temperatures, the expansion was proposed to occur earlier at lower

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latitudes and later cross the pCO_2 threshold at cooler, higher latitudes (e.g., Cerling et al., 1997). During the Pliocene, pCO_2 also decreased (Fedorov et al., 2013), but ecosystems at higher latitudes or elevations remained relatively stable, perhaps sufficiently past the pCO_2 threshold. In contrast, the tropical ecosystems adjacent to the Indian Ocean seem to be more sensitive, suggesting pCO_2 is likely not the primary driver of the Pliocene C_4 expansion.

Other possible triggers for C_4 expansion (aridity, seasonality, flood, droughts, and fires) depend on precipitation patterns. In the modern era, a complex interplay of multiple modes of variability dictate the unique seasonal and multi-decadal precipitation patterns in the regions surrounding the Indian Ocean. For example, rainfall on the Indian Peninsula follows quintessential monsoon behavior, but is also tied to the Inter-Tropical Convergence Zone (ITCZ), Walker Circulation, and the Indian Ocean Dipole (IOD; Wang et al., 2017). The biannual rains of the (semi)arid tropical East Africa are related to the ITCZ, monsoon winds, sea surface temperature (SST), and Walker Circulation (e.g., Williams and Funk, 2011; Tierney et al., 2015; Yang et al., 2015). Monsoon rains annually quench northern Australia, but the El Niño Southern Oscillation (ENSO), Walker Circulation, and the amount of Indonesian Throughflow (ITF) better explain the precipitation in other parts of Australia (Ummenhofer et al., 2009, 2011a, 2011b).

Since a single process cannot explain the precipitation patterns surrounding the Indian Ocean, the interplay of many processes likely changed during the Pliocene. Along with decreasing pCO_2, the ITF constricted (Cane and Molnar, 2001; Karas et al., 2009; Christensen et al., 2017) and SST gradients changed (Wara et al., 2005; Zhang et al., 2014; Burls and Fedorov, 2017), which would have modulated the monsoon dynamics and their interaction with the ITCZ, ENSO, IOD, and other modes of variability. Manifesting as increased aridity, seasonality, droughts, flooding, or fires, the changes in the hydroclimate variability over the Pliocene led to conditions more conducive to C_4 vegetation in certain tropical regions adjacent to the Indian Ocean.

5. Conclusion

Our study provides a piece of the puzzle in unraveling the complexities of C_4 expansion and adds nuance to the discussion of triggering mechanism. Although C_4 vegetation was established in the CMZ on the Indian Peninsula before the end of the Miocene, the results of this study show another significant expansion in the Pliocene (~3.5 to 1.5 Ma). The latter expansion is not observed in many records from the orographically-wet Himalaya emphasizing the spatial heterogeneities in C_4 vegetation response – even within the same monsoon system. However, other regions adjacent to the Indian Ocean, such as tropical East Africa and Northwest Australia, corroborate the observed expansion in the CMZ of the Indian Peninsula and show C_4 vegetation patterns sensitive to the changes in hydroclimate during the Pliocene.

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6. Appendices

Appendix A: Analytical Procedures - Inorganic Analyses of bulk major, trace, rare earth element concentrations

The samples we analyzed for major, trace, and rare earth element concentrations were originally collected for moisture and density (MAD) measurements onboard the JOIDES Resolution during IODP Expedition 353. Each sample was collected with a 2 cm diameter plastic syringe that fits into the top of a 10 cm^3 volume glass vial, allowing for the vial to be completely filled with sediment (Clemens et al., 2016). The samples were dried in a convective oven at 105°C ± 5°C for 24 hours (Clemens et al., 2016). The remaining sample preparation, digestions, and analyses were conducted at Boston University and a detailed description of the analytical geochemical procedures are presented in Dunlea et al. (2015). In summary here, sediment samples were hand-powdered with an agate mortar and pestle. For major and certain trace elements, sample powders were digested by flux fusion (Murray et al., 2000) and analyzed by inductively coupled plasma-emission spectrometry (ICP-ES). For analysis of additional trace and rare earth elements, sample powders were dissolved in a heated acid cocktail (HNO_3, HCl, and HF, with later additions of HNO_3 and H_2O_2 after samples were dried down) under clean-lab conditions and analyzed by inductively couple plasma-mass spectrometry (ICP-MS). Three separate digestions of a matrix-matched in-house sediment standard were analyzed with each batch and determined precision [(standard deviation)/(average) x 100] was ~2% of the measured value for each element. The international Standard Reference Material BHVO-2 was analyzed as an unknown with each batch and results were consistently found to be accurate within precision for each element.

Appendix B: Analytical Procedures - Carbon and nitrogen content and isotopes

Analyses of the abundance of total carbon (TC), total inorganic carbon (TIC), total organic carbon (TOC), nitrogen (N), the $\delta^{13}C$ of the TOC, and the $\delta^{15}N$ of the N component were performed at Woods Hole Oceanographic Institution and methods are described in Whiteside et al. (2011). In brief here, samples for TOC were weighed into tared silver boats and then acidified to remove carbonates in a closed desiccator for 3 days at 60-65°C over concentrated hydrochloric acid. All samples were flash combusted in a Costech 4010 Elemental Analyzer coupled via a Finnigan-MAT Conflo-II interface to a Thermo DeltaVPlus isotope ratio mass spectrometer. Data were recorded and integrated using the Isodat software package. Post-run calculations were performed for blank corrections, quantifications, and final calibrations.

Appendix C: Analytical Procedures - Compound specific biomarkers abundances and isotopes

The analyses of compound-specific biomarkers were performed at Brown University (e.g., Daniels et al., 2017). Samples were freeze-dried and lipids were extracted from 3.5 to 4.5 g of sediment using a Dionex 350 Accelerated Solvent Extractor (ASE) with dichloromethane:methanol (9:1 v/v). The fatty acids in the total lipid extract were separated from the neutral lipids using aminopropyl silica gel chromatography, eluting with a dichloromethane:isopropanol solution followed by ether with 5% acetic acid.

The fatty acids were methalyated to form fatty acid methyl ester (FAME) by dissolving dried down acid fraction in in ~0.3 mL of toluene and ~1mL of 5:95 acetyl chloride:methanol. Nitrogen replaced the headspace in the vial before they were capped tightly and heated at 60°C for 12 hours. Once the reaction was complete, the FAMEs were separated from the water by-products formed during the methylation reaction. Sample received ~1mL of synthetic saline solution (50g NaCl/L of double-distilled water) and ~1mL of hexane, were vigorously shaken, and then allowed to rest until the hexane separated from the water. The hexane fraction was pipetted into a new vial, avoiding the water. Another ~1mL of hexane was added to the sample, shaken, and pipetted into the new vial. To clean the solution and isolate the fatty acids, samples were run through a second silica gel column, eluting with hexane to remove unwanted acids and then DCM to acquire the clean FAME fraction.

The neutral lipid fraction of the total lipid extract was further separated into the n-alkane fraction (N1), ketone/sterol fraction (N2, alkenones), and the alcohol fraction (N3, GDGTs and diols) eluting with hexane, DCM, and methanol, respectively. The FAME, alkanes (N1), and alkenone (N2) fractions were analyzed on an Agilent 6890 gas chromatograph with a flame ionization detector (GC-FID). To quantify alkane (N1) abundances, a hexamethylbenzene internal standard was analyzed with each sample. To quantify alkenone (N2) abundances, an 18-pentatriacontane standard was used. Sample blanks were analyzed with every batch.

The isotope ratios of the FAME fraction ($\delta D_{n\text{-acid}}$ and $\delta^{13}C_{n\text{-acid}}$) were measured on a Thermo Finnigan Delta + XL isotope ratio mass spectrometer with a HP 6890 gas chromatograph and a high-temperature pyrolysis reactor for sample introduction. For δD, three injections of each sample were analyzed and two injections of each sample were analyzed for $\delta^{13}C$. Between every six injections, a standard mixture containing C_{16}, C_{18}, C_{22}, C_{26}, and C_{28} n-acids was analyzed to monitor instrument accuracy and precision. Analytical uncertainty was calculated by [standard deviation/average] of the injections and is typically less than 3% for δD and less than 1% for $\delta^{13}C$. The standard deviations are reported in Table S3. For every instrument run, samples were analyzed in random order.

Data Availability

Data is included in the supplementary tables and will be made publically available in the NSF funded EarthChem database (https://www.earthchem.org/) maintained by the Interdisciplinary Earth Data Alliance (IEDA) at the Lamont-Doherty Earth Observatory of Columbia University.

Author Contribution

A.G.D. participated in the IODP Expedition 353 sampling party, performed the geochemical analyses, and lead writing and revisions of the manuscript. L.G. participated on IODP Expedition 353, was involved in the project's conceptualization, sample acquisition, and provided supervision. Y.H. also participated on IODP Expedition 353, was

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involved in the project's conceptualization, provided resources and funding acquisition, advised on the methodology, supervision, and aided in interpretation of data.

Competing Interests

The authors declare that they have no conflict of interest.

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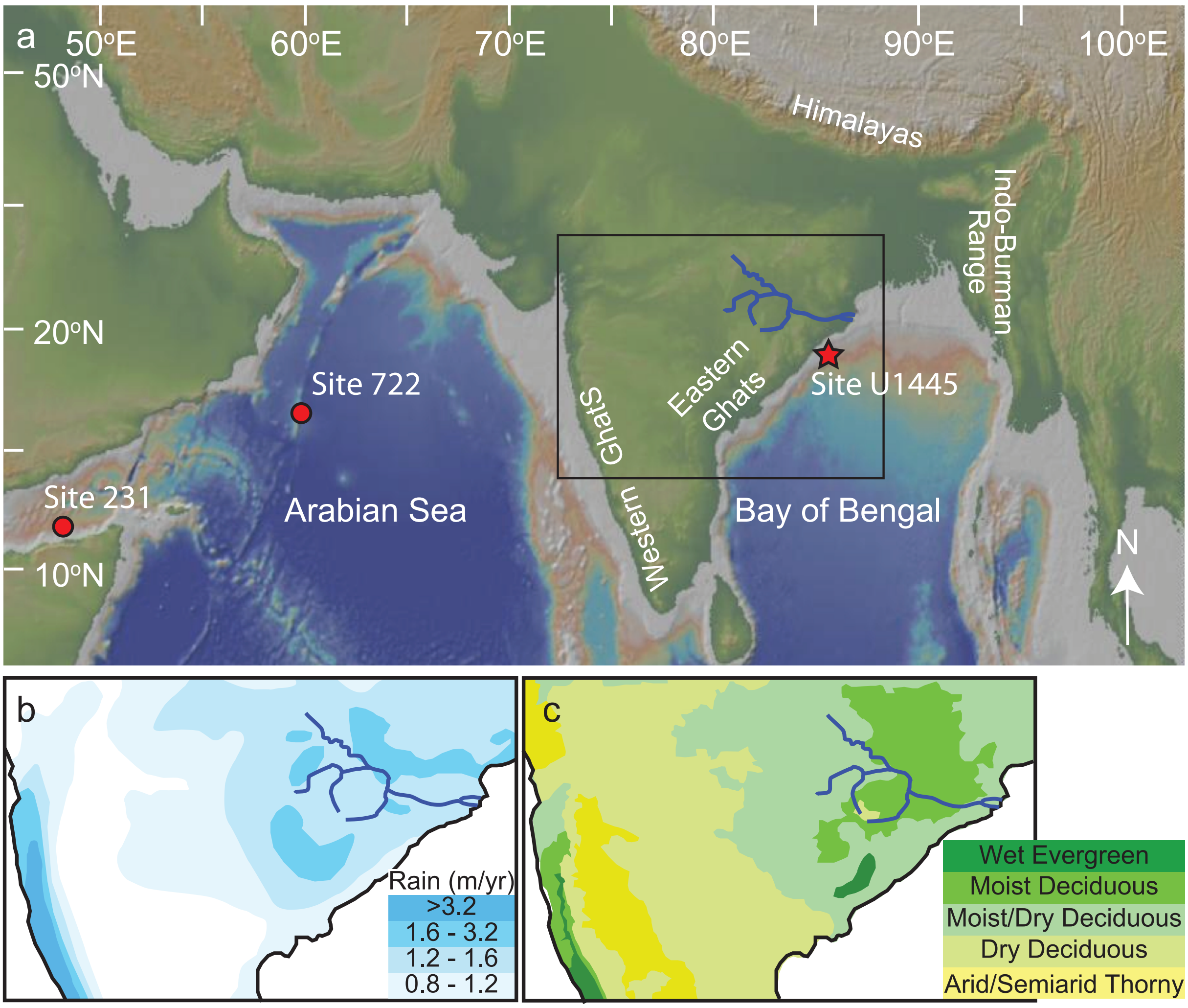
Figure 1. (a) Location of IODP Site U1445 in the Bay of Bengal (red star). Site 231 in the Gulf of Aden and Site 722 in the Arabian Sea are plotted for reference (red dots). Topography and bathymetry are represented in the background map. The Mahanadi River and main tributaries are traced in dark blue and the region outlined by the box is zoomed-in for Figures 1a and 1c, which are modified from Ponton et al. (2012). (b) average annual rainfall (m/year) and (c) natural ecosystems in the region including the Mahanadi River drainage basin.



415 **Figure 2. Analyses of 57 samples from Site U1445 in the Bay of Bengal. Black and white dots are pairs of samples from relatively dark and light layers, respectively, at a similar depth. Triangles are samples not in pairs. Black curves are a 9-point moving average of all samples. Black text labels the data and brown text is our interpretations. (a) total organic carbon (wt. %). The grey line represents shipboard measurements (Clemens et al., 2016). (b) carbon isotope values of C_{30} fatty acids from leaf waxes ($\delta^{13}C_{FA}$, per mil), (c) hydrogen isotope values of C_{30} fatty acids from leaf waxes (δD_{FA}, per mil). The yellow horizontal band**
420 **highlights the Pliocene epoch.**

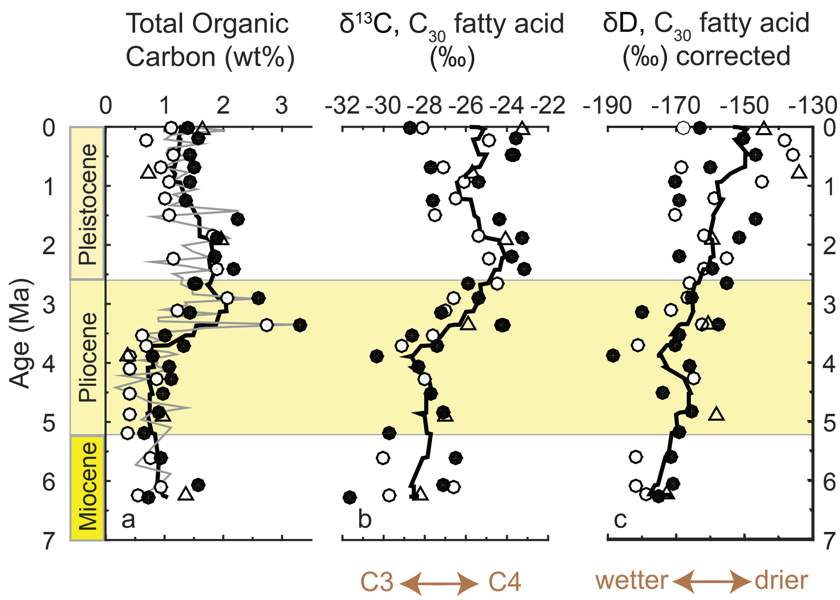


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Figure 3. (a) $\delta^{13}C$ of C_{33}-alkanes from Siwalik paleosols in Northern Pakistan (white dots) and from sediment at Site 717 in the Bengal Fan (black dots; Freeman and Colarusso, 2001). (b) $\delta^{13}C$ of C_{31}-alkanes at Site 722 in the Arabian Sea (Huang et al., 2007), which integrates vegetation variability from north and east of the Arabian Sea. (c) $\delta^{13}C$ of C_{28}-fatty acids at Site 231 in the Gulf of Aden, which records vegetation in East Africa (Feakins et al., 2013; Liddy et al., 2016). (d) $\delta^{13}C$ of C_{30}-fatty acid at Site U1445 in the Bay of Bengal, which records vegetation from the Mahanadi basin on the Indian Peninsula (this study). (e) $\delta^{13}C$ of C_{33}-alkanes from northwest Australia (Andrae et al., 2018). Brown arrows point out the late-Miocene C_4 expansion in Northern Pakistan, Arabian Sea, and Gulf of Aden as well as the Pliocene C_4 expansion in northeast Africa and the Indian Peninsula.