Lateral variations in vegetation in the Himalaya since the Miocene and implications for climate evolution

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Abstract

The Himalaya has a major influence on global and regional climate, in particular on the Asian monsoon system. The foreland basin of the Himalaya contains a record of tectonics and paleoclimate since the Miocene. Previous work on the evolution of vegetation and climate has focused on the central and western Himalaya, where a shift in vegetation has been observed at ~7 Ma and linked to increased seasonality, but the climatic evolution of the eastern part of the orogen is less well understood. In order to track vegetation as a marker of monsoon intensity and seasonality, we analyzed $\delta^{13}C$ and $\delta^{18}O$ values of soil carbonate and associated $\delta^{13}C$ values of bulk organic carbon from previously dated sedimentary sections exposing the syn-orogenic detrital Dharamsala and Siwalik Groups in the west, and, for the first time, the Siwalik Group in the east of the Himalayan foreland basin. Sedimentary records span from 20 to 1 Myr in the west (Joginder Nagar, Jawalamukhi, and Haripur Kolar sections) and from 13 to 1 Myr in the east

(Kameng section), respectively. The presence of soil carbonate in the west and its absence in the east is a first indication of long-term lateral climatic variation, as soil carbonate requires seasonally arid conditions to develop. $\delta^{13}C$ values in soil carbonate show a shift from around -10 ‰ to -2 ‰ at ~7 Ma in the west, which is confirmed by $\delta^{13}C$ analyses on bulk organic carbon that show a shift from around -23 ‰ to -19 ‰ at the same time. Such a shift in isotopic values is likely to be associated with a change from C3 to C4 vegetation. In contrast, $\delta^{13}C$ values of bulk organic carbon remain at ~-23 ‰ in the east. Thus, our data show that the current east-west variation in climate was established at 7 Ma. We propose that the regional change towards a more seasonal climate in the west is linked to a decrease of the influence of the Westerlies, delivering less winter precipitation to the western Himalaya, while the east remained annually humid due to its proximity to the monsoonal moisture source.

1. Introduction

The Himalayan belt has a major influence on global and regional climate, by acting as an orographic barrier for air masses and humidity (Boos and Kuang, 2010; Molnar et al., 2010). Modern climate shows significant east-west variation in the Himalaya; both mean-annual and winter precipitation on the plains and foothills are higher in the east, while the west is characterized by more pronounced winter aridity (Figure 1; Bookhagen and Burbank, 2006; 2010). This variation is due to the two major atmospheric circulation systems influencing the climate of the Himalayan region: the Indian Summer Monsoon (ISM) and the Westerlies (Kotlia et al., 2015). The ISM takes up moisture in the Bay of Bengal and transports it towards the Himalaya during the northern-hemisphere summer months (e.g., Molnar et al., 2010), whereas the Westerlies bring moisture from the Mediterranean, Black and Caspian Seas and are most efficient in winter (Benn and Owen, 1998; Cannon et al., 2015). Generally, the influence of the Westerlies is greater in the western part of the Himalayan region (Cannon et al., 2015; Caves et al., 2015; Kotlia et al., 2015). The proximity to the moisture source in the Bay of Bengal, makes the eastern Himalaya very humid (Bookhagen et al., 2010).

These lateral variations in modern climate are linked to vegetation patterns, in particular the relative importance of C3 versus C4 plants. C3 plants are favored in a cool and humid climate, whereas C4 plants prefer intense light, warm and water-stressed conditions (Ehleringer, 1988). An additional factor that possibly influences the evolution of C4 plants is growing season temperature and precipitation, favoring C4 plants in drier periods (Cotton et al., 2016). The distinct stable carbon-isotopic signature of C3 versus C4 vegetation allows paleo-vegetation to be tracked from the sedimentary record: pure C3 vegetation has $\delta^{13}C_{org}$ values between -22 ‰ and -30 ‰, whereas $\delta^{13}C_{org}$ values of C4 plants range from -10 ‰ to -14 ‰ (Cerling et al., 1997). The lateral variations in modern climate and vegetation are expressed by the signature of modern organic carbon transported in the foreland (Galy et al., 2008a; Fig. 1): sediments sampled from Himalayan tributaries at the mountain front have $\delta^{13}C_{org}$ values around -25 ‰, indicating dominance of C3 plants at higher elevations within the mountain belt. These values remain stable within the eastern Brahmaputra catchment, whereas they increase to values around -22 ‰ in the Western Ganga catchment (Fig. 1), implying laterally varying vegetation (from C4 in the west to C3 in the east) in the floodplain.

An important question is when and why the modern spatial patterns in climate and vegetation were established in the Himalayan foreland. The onset of the ISM is dated back to at least the middle Miocene (Dettman et al., 2001) and possibly the Late Eocene (Licht et al., 2014). Likewise, the Westerlies have been argued to influence Asian climate since the Eocene (Caves et al., 2015). However, the evolution of regional climate and vegetation patterns will depend on the relative strength of these two systems through time, which remains largely unknown.

The foreland basin of the Himalaya contains a sedimentary record of vegetation and paleoclimate since Miocene times, within the continental detrital pre-Siwalik and Siwalik Groups. The record of spatial and temporal variations in vegetation holds information on climate evolution, in particular patterns of atmospheric circulation, seasonality and the origin and transport of humidity (Hoorn et al., 2000; Sanyal et al., 2004; Gupta 2010, amongst others). Carbon and oxygen isotopic compositions of soil carbonates and soil organic matter from pre-

Siwalik and Siwalik sediments in Nepal, Northwest India and Pakistan have been used to reconstruct changes in vegetation and climate during the Neogene (Quade et al., 1989, 1995a; Quade and Cerling, 1995; Sanyal et al., 2010; Singh et al., 2013). These records consistently show a shift in $\delta^{13}C$ values at ~7 Ma, which has been interpreted as a change from C3 to C4 vegetation, which was initially interpreted to be related to an intensification of the ISM (Quade et al., 1989). Steinke et al. (2010) suggest that this change was rather linked to an increase in aridity, and therefore a weakening of the ISM. A similar shift has also been recorded in the distal Himalayan-derived sediments of the Bay of Bengal (France-Lanord and Derry, 1994). It has been argued that the late-Miocene expansion of C4 plants is a global phenomenon due to a decrease in atmospheric pCO_2 (Cerling et al., 1997), global cooling and/or increased aridity (Herbert et al., 2016). Others, however, argue that pCO_2 was already at a level favorable for C4 plants during Oligocene times (Pagani et al., 2005; Beerling and Royer, 2011) and that the spread of C4 plants during the late Miocene should therefore have other, more regional triggers.

The focus of previous studies on Himalayan climate and vegetation records has been entirely on the western and central Himalayan foreland; no climate and vegetation data are available east of Nepal. In order to obtain better spatial insight into the evolution of the monsoon climate, precipitation patterns and the expansion of C4 plants along strike in the Himalaya, we present and compare new $\delta^{13}C$ and $\delta^{18}O$ data of pedogenic carbonate and organic matter from the northwestern and the poorly studied eastern Himalayan foreland basin. Lateral variations in the evolution of the vegetation yield further insight into how and under what climatic conditions C4 plants developed or not, suggesting that regional influences play a major role.

2. Setting

Neogene Himalayan foreland-basin sediments are composed of the fluvial Dharamsala Group of Late Oligocene and early Miocene age (Burbank et al., 1996), and the Siwalik Group deposited since the early Miocene. The Dharamsala rocks consist of continental fluvial, lacustrine or deltaic sediments, and contain fine-to medium grained sandstones, siltstones and overbank mudstones with soil carbonate nodules (Raiverman and Seshavataram, 1983). The sediments of the Siwalik

Group are exposed nearly continuously along the front of the Himalayan range, with only minor age variation along strike (Burbank et al., 1996). They were deposited in the foreland before being incorporated in the foothills due to southward propagation of deformation and onset of motion on the Main Frontal Thrust (MFT).

The Siwalik Group shows an overall coarsening- and thickening-upward trend, interpreted as recording increasingly proximal deposition (DeCelles et al., 1998), and is divided into the Lower, Middle and Upper Siwaliks (LS, MS, US). The LS were deposited by high-sinuosity streams (Nakayama and Ulak, 1999). The Middle Siwaliks (MS) are characterized by thickly bedded sandstones, which are medium- to coarse-grained and often rich in mica. The MS represent a depositional environment of large braided rivers. The Upper Siwaliks (US) consist of beds of conglomerates alternating with sandstone beds, deposited by gravelly braided rivers. Paleosols are developed throughout most of the Siwalik sections, with lateral and temporal variations in abundance: they are more abundant in the LS and in the west. Paleosols are characterized in western and central Himalayan sections by the presence of soil-carbonate nodules.

We sampled three sections exposing Dharamsala and Siwalik deposits in the western Himalaya; the Joginder Nagar (JN), Jawalamukhi (JW) and Haripur Kolar (HK) sections in Himachal Pradesh, and one Siwalik section in the eastern Himalaya; the Kameng River (KM) section in Arunachal Pradesh (Figures 1, 2). All sections have previously been dated by magnetostratigraphy (Meigs et al., 1995; Sangode et al., 1996; White et al., 2001; Chirouze et al., 2012). They span a time range of 20-1 Ma in the west and 13-1 Ma in the east. In the western sections, we collected paleosols and associated carbonate nodules, as well as fine-grained mudstone in zones without well-developed paleosols. Carbonate nodules are lacking in the Kameng section (Figure 2); therefore only mudstones, where possible from paleosols, were sampled. Additionally, modern river mud was sampled from riverbanks in proximity to the sections in both the west and the east. Modern river samples in the west (Beas River and Jner Khad River) were collected at an elevation of ~640 m within the sections, whereas the modern Kameng River sample was collected at an elevation of ~100 m downstream of the Siwalik.

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3. Methods

$^{13}C/^{12}C$ and $^{18}O/^{16}O$ ratios (expressed as $\delta^{13}C$ and $\delta^{18}O$ values respectively) of soil carbonate nodules were determined using a multiflow analyser linked to an Isoprime 100 continuous flow mass spectrometer at the Lancaster University, UK. Approximately 600-700 µg of sample powder was drilled from each carbonate nodule and digested online at 90°C with dehydrated phosphoric acid in a He-flushed exetainer. Product CO_2 was analyzed for $\delta^{13}C_{CO2}$ and $\delta^{18}O_{CO2}$ and corrected against VPDB and VSMOW, respectively, using within-run analyses of international standards NBS18, LSVEC and CO-1. Within-run standard replication (1 σ) was <0.1 ‰ for both C and O isotope ratios. Sample replication based on separate drill aliquots of powder from the same carbonate nodule was <0.1‰ for $\delta^{13}C$ and <0.2 ‰ for $\delta^{18}O$ (1 σ).

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$^{13}C/^{12}C$ ratios (expressed as $\delta^{13}C$ values) of bulk organic matter were determined using an Elementar Vario Micro elemental analyser linked to a VisION continuous flow mass spectrometer at the University of Lancaster. The carbonate content of each sample was removed by acid digestion using 1M ultrapure HCl and the resultant sample washed repeatedly using de-ionised water and centrifugation. Approximately 10 mg of each prepared sample was combusted within tin capsules at 960 °C to yield CO_2 for determination of $\delta^{13}C_{org}$. $\delta^{13}C$ values were corrected against VPDB using internal reference materials calibrated to international standards. Within-run $\delta^{13}C$ replication (1 σ) was <0.2 ‰ for standards and <0.25 ‰ for samples.

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4. Results

The three sections in Himachal Pradesh (Western Himalaya) provide a continuous age record over the past 20 Ma (Figure 3). Prior to ~7 Ma, $\delta^{13}C$ values of soil carbonate ($\delta^{13}C_{soil\ carb.}$) range between -8 ‰ and -13 ‰, whereas at ~7 Ma a shift towards more positive $\delta^{13}C$ values, ranging from +2 ‰ to -8 ‰, is observed (Figure 3; Appendix 1). $\delta^{18}O$ values range mostly from -11 ‰ to -4 ‰, except in the older part of the sections, where some values are as low as -14 ‰. A weak trend towards more positive $\delta^{18}O$ values over time is observed. As noted above, soil carbonate was not present in the eastern Himalayan section.

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We additionally measured a continuous record of $\delta^{13}C$ in organic carbon ($\delta^{13}C_{org}$) in both the western and eastern sections (Figure 4; Appendix 1). In the west, a clear shift towards more positive $\delta^{13}C_{org}$ values is observed at ~7 Ma, synchronous with the $\delta^{13}C_{soil\ carb}$. Before 7 Ma, $\delta^{13}C_{org}$ values range between -23 ‰ and -27 ‰, while values are less negative, from -18 ‰ to -23 ‰, after 7 Ma. In the east, in contrast, $\delta^{13}C_{org}$ values remain constant between -29 ‰ and -23 ‰ since the middle Miocene and no shift towards more positive values is observed. Organic matter from modern river sediments show $\delta^{13}C$ values of approximately -26 ‰ and -23.5 ‰ in the west and in the east, respectively (Figure 4). Total Organic Carbon content (TOC) in the western samples is mostly <0.5 % although samples from the JN can have up to 15% TOC (Figure 5). TOC of most Kameng samples is also <0.5 %, with some samples showing values up to 3%. There is no correlation between $\delta^{13}C_{org}$ values and TOC (Appendix 2).

5. Discussion

5.1. Modern river sediments and vegetation

The modern Ganga/Brahmaputra floodplain is widely used for agriculture and is therefore predominantly covered in C3 plants such as rice crops, and in the east by tea plantations (Blasco et al., 1996). This is not reflected by the organic carbon transported in the Ganga and Brahmaputra Rivers: $\delta^{13}C_{org}$ values of ~-21.9 ‰ in the modern Ganga floodplain are representative of a mixture of C4 and C3 plants, suggesting the presence of C4 plants in the west, whereas the modern Brahmaputra River carries organic carbon with $\delta^{13}C_{org}$ values of -23.0 ‰ in the range of C3 plants (Galy et al., 2008a).

Measured $\delta^{13}C_{org}$ values of modern river muds in both the west (Beas and Jner Khad River) and the east (Kameng River) are in the range of C3 plants. However, these modern river sediments were not collected in the floodplain but at the mountain front, where they will contain detrital organic carbon of C3 plants transported from higher elevations (Dobremez, 1978) and/or fossil organic carbon from Himalayan formations (Galy et al., 2008a), which both have a more negative isotopic signal. Dharamsala and Siwalik sediments were deposited further into the floodplain, and hence should carry a signal of floodplain vegetation.

5.2. Possible factors influencing the isotopic signal

Earlier studies (Quade et al., 1995a; Quade and Cerling, 1995; Singh et al., 2007; Sanyal et al., 2010) measured $\delta^{13}C$ on soil carbonate nodules, which can only be found in the western and central Himalayan Siwalik sections, and have consistently shown a change towards more positive values at ~7 Ma from Pakistan to Nepal (Quade et al., 1995a; Quade and Cerling, 1995). This change was interpreted in terms of a shift in vegetation, from initially dominated by C3 plants to containing C4 species. In order to compare the western and the eastern Himalaya in this study, we rely on $\delta^{13}C_{org}$ of bulk organic matter, as soil carbonate nodules are absent in the east. In both the western and the eastern sections, $\delta^{13}C_{org}$ values range between -23 ‰ and -29 ‰ before 7 Ma, indicating vegetation dominated by C3 plants. After 7 Ma, $\delta^{13}C_{org}$ in the western sections demonstrates an isotopic shift to values enriched in ^{13}C (~-19 ‰), suggesting that a component of the organic matter comprises C4 species. Carbonate nodules from western Himalayan sections analyzed in this study show a trend similar to $\delta^{13}C_{org}$, demonstrating a shift from C3-dominated vegetation composition, to an increasing proportion of C4 species in the younger sections. In the east, in contrast, $\delta^{13}C_{org}$ values stay in the range of C3 plants throughout the sedimentary succession (Figure 4).

In modern soils, carbonate precipitates in equilibrium with soil CO_2 (Cerling et al., 1989) following an isotopic enrichment in ^{13}C of 10.36 ‰, (Cerling et al., 1989). Diffusional effects cause soil-respired CO_2 to be further enriched in ^{13}C by 4.4 ‰. The total fractionation between soil organic matter and soil carbonate is ~14 ‰, at 25°C to ~17 ‰ at 0°C (Cerling et al., 1989). $\delta^{13}C_{soil\ carb.}$ values show a greater (~ +10‰) shift towards more positive values after 7 Ma than the $\delta^{13}C_{org}$ values (~ + 6 ‰; Figure 6). This discrepancy in the absolute value of the isotopic shift to signatures more enriched in ^{13}C likely reflects the nature of carbonate nodule production and organic matter source. Whereas soil carbonates reflect only the soil CO_2 characteristics and temperature during formation, the $\delta^{13}C_{org}$ values are more susceptible to bias by inherited and transported material.

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Although $\delta^{13}C_{org}$ is usually interpreted to represent isotopic values of vegetation in the floodplain at the time of sediment deposition, it can potentially be biased by several factors, such as input of (likely C3 plant dominated) organic carbon from high elevations (Dobremez, 1978) and/or input of fossil organic carbon. The amount of fossil organic carbon present in Himalayan river sediments was estimated using the radiocarbon content of total organic carbon (TOC) of modern suspended and bedload sediments (Galy et al., 2008a; 2008b). Galy et al. (2008b) estimated the total amount of fossil organic carbon transported in the Ganga and Brahmaputra Rivers between 0.02 and 0.03 %. Sediments of the sampled sections mostly have TOC values >0.1 % (Figure 5) suggesting that the proportion of fossil carbon should be low, <30 % at most, if the modern amounts of transported fossil organic carbon can be extrapolated to the past. $\delta^{13}C_{org}$ of fossil organic carbon from Himalayan source rocks varies from -28 to -14.6‰ (Galy et al., 2008a); it is therefore unclear what the effect of varying proportions of fossil organic carbon on the observed $\delta^{13}C_{org}$ signal would be. However, we have no reason to assume the influence of fossil carbon to be very different from east to west and our data suggest this influence to be rather constant over time (see below). It is therefore unlikely that the spatial and temporal variations in $\delta^{13}C_{org}$ values could be explained by variable fossil organic carbon content.

Additionally, the efficiency of oxidation of organic carbon and hence the replacement of inherited carbon by floodplain carbon can influence the $\delta^{13}C_{org}$ signal in modern river sediments (Galy et al., 2008a; 2011). Organic-carbon oxidation varies between the Ganges and Brahmaputra foreland basin, being more efficient in the Ganges floodplain due to different hydrological settings: the Ganges is a meandering river, whereas the Brahmaputra is a braided river with a narrower floodplain (Galy et al., 2008a). Oxidation of inherited organic carbon from vegetation at higher altitudes, hence with a C3 signal, is efficient in the Ganges floodplain (Galy et al., 2008a), as shown by the proportion of C4 organic matter increasing downstream in the Ganges floodplain (Figure 1). This pattern is not present in the Brahmaputra floodplain, suggesting that the influence of inherited carbon could therefore be greater in the Brahmaputra. Even though paleosols are less developed in the east, high TOC values indicate the presence of organic matter acquired from surface organic litter during pedogenesis (Figure 5). Degradation of organic matter in soils could have an influence on the $\delta^{13}C_{org}$: detrital organic matter has a ~1-2 ‰ more negative $\delta^{13}C$ signal than soil organic matter (von Fischer and Tieszen, 1995 and references therein). The more negative $\delta^{13}C$ values in the Kameng section could therefore be explained by the presence of more detrital organic matter. Differences in floodplain dynamics during transport and a greater influence of inherited organic carbon could possibly bias the signal but are unlikely to cancel out the entire C4 signal in the eastern Himalaya.

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Further information on different organic carbon sources could potentially be derived from lipid biomarker analysis (i.e. compound-specific C- and H-isotope analysis; Freeman et al. 2001; Sachse et al., 2012). We extracted n-alkanes from samples of both the western and the eastern sections (see Appendix 3), but unfortunately n-alkane preservation was generally low in the sediments. Moreover, evidence of diagenetic overprinting was found in sediments with a sufficient concentration, evidenced by an absence of the predominance of odd carbon numbered n-alkane chain lengths (expressed as the carbon preference index, CPI), which is prevalent in modern plant and sediment samples. In modern plant material and immature sediments, CPI values are generally significantly >3 and up to 20, whereas we found values around 1 in the Kameng samples, indicating diagenetic overprinting or addition of fossil carbon at the time of sedimentation. As a result of this, compound-specific hydrogen and carbon stable isotopic values would likely have been altered towards less negative values (Radke et al., 2005). While diagenetic overprinting could also have affected bulk $\delta^{13}C_{org}$ values by homogenizing the isotopic signal (Cerling et al., 1984; Bera et al., 2010), our CPI data remained uniform at values around 1 from the base of the section until ca. 2 Ma, indicating a similar degree of overprinting in these samples. Since we did not find any change in bulk $\delta^{13}C_{org}$ values corresponding to changes in CPI values, we argue that any potential overprinting affected all samples equally and as such relative changes can still be interpreted from bulk $\delta^{13}C_{org}$ values. In addition, the samples presented in Figure 3 are in the same isotopic range as modern soil carbonate nodules, indicating that diagenesis is unlikely to have influenced the isotopic values of the sedimentary samples.

The isotopic values of pedogenic carbonate are in equilibrium with soil CO_2 derived from irreversible oxidation of organic matter in a diffusion-controlled soil system at different temperatures. The isotopic equilibrium factor is dependent on temperature, hence if pedogenic carbonate precipitates in equilibrium with soil CO_2, $\delta^{13}C_{soil\ carb}$ should be enriched by ~14 ‰ at 25 °C and by 17 ‰ at 0 °C, respectively (Cerling et al. 1984, 1989). Carbonate nodules and their corresponding organic matter of the Dharamsala and Siwalik paleosols plot mostly at temperatures between 0 and 25°C and are therefore not isotopically altered by diagenesis (Figure 6), but rather formed within this soil temperature range. Samples below the 0°C line could reflect an inconsistency in the enrichment of ^{13}C in soil carbonate nodules compared to the co-existing organic matter. Most of the samples plotting under the 0°C line show a strong C4 signal, and the offset between $\delta^{13}C_{org}$ and $\delta^{13}C_{soil\ carb.}$ in these samples is greater than expected from isotopic equilibrium considerations. A possible explanation for this enhanced offset could be that the carbonate nodule formed in a sediment body (e.g. another paleosol horizon) that was separate from the parent organic matter. Alternatively, organic matter may be more influenced by inherited organic matter from C3 vegetation, whereas the $\delta^{13}C_{soil\ carb}$ would more directly represent the local vegetation cover. However, this inconsistency does not reflect a diagenetic overprint, as it is found in the youngest samples, where diagenesis is least likely to occur.

$\delta^{18}O_{soil\ carb}$. values of the three western sections show a slight change towards more positive values (Figure 3), comparable to $\delta^{18}O$ values of the Surai Khola section in Nepal (Quade et al., 1995a). Only $\delta^{18}O_{soil\ carb.}$ values from Pakistan show a clear shift from values < -8 towards more positive values at ~8-6 Ma (Quade and Cerling 1995). In contrast, samples from this study already show $\delta^{18}O_{soil\ carb.}$ values > -8 before 7 Ma (Figure 3). A change in $\delta^{18}O_{soil\ carb.}$, which forms in-situ from soil water, can be associated with a change in either soil temperature ($\delta^{18}O_{soil\ carb.}$ being positively correlated with mean annual temperature; Cerling, 1984) and/or precipitation source; $\delta^{18}O$ values of precipitation of moisture transported from the Bay of Bengal are generally lighter (more negative) than $\delta^{18}O$ values of moisture transported by the Westerlies (Caves et al., 2015 and references therein). The isotopic change over time was measured on samples of three separate sections (Figure 2) at different longitudinal locations; therefore the isotopic signature from precipitation may be location specific rather than representing change over time. However, there is no clear shift in $\delta^{18}O_{soil\ carb.}$ values going from one section to another (Figure 3), suggesting this effect to be minimal. As all sediments were deposited in the foreland, the influences of any altitudinal effects (Dansgaard, 1961) can also be excluded.

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The coarsening-upward trend of sedimentary rocks throughout the sections reflects a change in depositional environment and location in the foreland basin, which varies from a distal floodplain for the Dharamsala and Lower Siwaliks to deposition closer to the mountain front in the Upper Siwaliks. At different locations in the foreland basin, the source of precipitation may vary from moisture influenced by the Westerlies, to moisture sourced from the ISM. More positive $\delta^{18}O_{soil\ carb.}$ values over time could therefore indicate an increasing influence of Westerlies with respect to ISM moisture sources, and/or a trend towards a warmer, drier climate, conducive to the growth of C4 vegetation.

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5.3. What caused the change of vegetation at ~7 Ma?

C3 and C4 plants grow in different environments and the $\delta^{13}C$ signal can therefore be used as an indirect climate indicator. Our data show that a change in vegetation occurred at ~7 Ma in the western Himalaya, but not in the east, where C3 plants have been dominant since the middle Miocene. As we have argued above, differences in floodplain setting (Galy et al., 2008a, 2011), while influencing the signal, cannot explain the observed lateral difference and neither can input of fossil organic carbon (Galy et al., 2008b). For this reason, there must be a remarkable lateral variation in the evolution of climate in the Himalayan region. The change at 7 Ma in the west and central Himalaya has been interpreted as resulting from a "stronger monsoon", characterized by greater seasonality (Quade et al., 1989, 1995a; Quade and Cerling, 1995). However, increased seasonality does not necessarily reflect higher amounts of monsoon precipitation; it could also indicate relatively less winter precipitation and thus a more arid (annual-average) climate (Molnar, 2005). C3 plants in the east indicate lower seasonality and higher annually averaged precipitation, consistent with modern precipitation patterns (Bookhagen and Burbank, 2010).

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The expansion of C4 plants in the west could therefore be a consequence of decreased winter precipitation, hence more seasonality associated with less (annually averaged) humidity, leading to a more arid climate. Overall, this difference in the $\delta^{13}C$ composition post-7 Ma is proposed to reflect water availability, with lower water availability in the west initiating a decline in C3 plants and a rise in C4 species (cf. Freeman and Colarusso, 2001). Dettman et al. (2001) likewise suggest a change in Indian summer monsoon characteristics and drying of the climate at 7.5 Ma. This scenario is supported by a change in $\delta^{18}O_{soil\ carb.}$ towards more positive values.

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Higher humidity in the east could be explained by the proximity to the Bay of Bengal, which is the major moisture source of precipitation in this area (Bookhagen et al., 2005). The western Himalaya is influenced by the Westerlies (Kotlia et al., 2015), which bring in winter precipitation. A decrease in the intensity of the Westerlies at 7 Ma would lead to more seasonality in the western floodplain, with drier periods in winter. An alternative explanation for a generally more arid climate in the western Himalaya could be a decrease of moisture transport from the Bay of Bengal and the Arabian Sea, possibly linked to a decrease in the intensity of the ISM. However, this would result in less seasonality, hence a less favorable climate for C4 plants. The spatially variable record of $\delta^{13}C_{org}$ values strongly suggests that the change in vegetation at 7 Ma did not occur simultaneously along the Himalayan foreland, indicating that the change is at least partly driven by regional factors rather than being linked only to a global change in atmospheric pCO_2. This supports the findings of Pagani et al. (2005) and Beerling and Royer (2011), who noted that atmospheric pCO_2 levels favoring C4 plants were already reached during the Oligocene. Other dry regions such as the Mediterranean have been dominated by C3 plants since the Miocene (Quade et al., 1994, 1995b), also indicating that the late-Miocene expansion of C4 plants was not a global phenomenon. Regionally dependent factors, such as differences in seasonality or humidity, have clearly played a role in determining Himalayan vegetation patterns through time. Lateral variations in vegetation suggest that there is a threshold somewhere along the Himalayan front, where the amount of (either annual or winter) precipitation becomes too large for C4 plants to spread.

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6. Conclusions

Stable carbon and oxygen isotopes were analyzed in carbonate nodules of the Joginder Nagar, Jawalamukhi and Haripur Kolar sections in the western Himalaya. $\delta^{13}C_{soilcarb.}$ values show a clear shift towards more positive values at 7 Ma, similar to the results of earlier studies in the western and central Himalaya. The lack of carbonate nodules in Siwalik sediments of the Kameng section, eastern Himalaya, is a first indicator that the lateral environmental and climatic differences in the modern Himalaya are representative of long-term climatic patterns. In order to directly compare the western and eastern sections, stable carbon isotopes on organic matter were analyzed and show a clear spatial difference. In the west, $\delta^{13}C_{org}$ values shift towards more positive values at 7 Ma, consistent with the results on carbonate nodules, whereas they remain constant over the last 13 Ma in the east. The $\delta^{13}C$ of organic matter reflects the evolution of vegetation, with the development of C4 plants in the west and an environment that remains favorable for C3 plants in the east. Such variations in vegetation imply differences in climate, which became more seasonal and overall drier in the west at 7 Ma. The eastern Himalaya is more proximal to the main moisture source for precipitation (the Bay of Bengal); therefore, even though climate may have varied, it remained less seasonal and more humid, inhibiting the evolution of C4 plants. Therefore, the change in climate in the west and the onset of lateral variation is most likely caused by a change in strength of atmospheric circulation, such as a weakening of the influence of the Westerlies. These findings suggest that the late-Miocene expansion of C4 vegetation does not depend solely on atmospheric pCO_2 but also on regional changes in aridity and seasonality. Newly developed methods, such as clumped isotopes or stable isotopes on compound-specific organic carbon, even though unsuccessful in this study, could provide further insight into the climatic evolution and the development of C4 vegetation, both globally and regionally in the Himalayan region. This study has provided the first paleo-climate and -vegetation data from the eastern Himalaya; however, more such studies are needed to refine our understanding of the evolution of climate and vegetation in this area.

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Figures

Figure 1: Map of the Himalayan region, with $\delta^{13}C$ of modern river organic carbon from Galy et al., 2008a. The Himalayan range is indicated schematically in grey. Sections are indicated in red: JW: Jawalamukhi; JN: Joginder Nagar; HK: Haripur Kolar; KM: Kameng. Lower plot shows comparison of modern annual precipitation data (TRMM) in proximity to the sampled sedimentary sections in the west and east.

Figure 2: Stratigraphy of sections of the Dharamsala and Siwalik Groups in the west (A) and in the east (B), with field photos showing sedimentological characteristics of different sub-groups.

Figure 3: A: $\delta^{13}C_{soil\ carb.}$ and $\delta^{18}O$ values of soil carbonate in the western Himalaya. Different symbols indicate the different sections (HK: Haripur Kolar; JW: Jawalamukhi; JN: Joginder Nagar). B: $\delta^{13}C_{soil\ carb.}$ and $\delta^{18}O$ values of soil carbonate in Pakistan from Quade and Cerling (1995).

Figure 4: $\delta^{13}C_{org}$ of bulk organic carbon in the western (HK: Haripur Kolar; JW: Jawalamukhi; JN Joginder Nagar) and the eastern (KM: Kameng) Himalayan sections. Light and dark grey shaded bars indicate $\delta^{13}C_{org}$ values characteristic of C3 and C4 plants, respectively (Cerling et al., 1997).

Figure 5: Total organic carbon content (TOC) vs age [Ma], zoomed in to values below 0.5 % on the left and values above 0.5 % on the right. TOC values above 0.07% indicate dominant biogenic C_{org}, from soil organic matter and floodplain vegetation, rather than detrital and fossil C_{org} (Galy et al., 2008b).

Figure 6: $\delta^{13}C_{soil\ carb.}$ of soil carbonate nodules vs. $\delta^{13}C_{org}$ of co-existing organic matter of the western Dharamsala and Siwalik Group sections. Solid and dashed lines represent isotopic values of pedogenic carbonate in isotopic equilibrium with the soil CO_2 derived from irreversible oxidation of organic matter in a diffusion controlled soil system at different temperatures (Cerling et al. 1989).

Appendices:

Appendix 1: Sample overview and results of $\delta^{13}C_{org}$, $\delta^{13}C_{soil\ carb.}$ and $\delta^{18}O_{soil\ carb.}$

Appendix 2: Total organic carbon (TOC) vs. $\delta^{13}C_{org}$ in the western (a) and eastern (b) sections, respectively.

Appendix 3: n-alkane analysis on the Kameng river section