Research Article

(U-Th)/He Thermochronology of the Indus Group, Ladakh, Northwest India: Is Neogene Cooling a Continental-Scale Thermal Event in the India-Asia Collision Zone?

G. Bhattacharya, D. M. Robinson, and D. A. Orme

1Department of Geological Sciences and the Center for Sedimentary Basin Studies, The University of Alabama, Tuscaloosa, AL 35487, USA
2Department of Earth Sciences, Montana State University, Bozeman Montana 59717, USA

Correspondence should be addressed to G. Bhattacharya; gbhattacharya1@crimson.ua.edu

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1. Introduction

The India-Asia continental collision zone archives a sedimentary record of the tectonic, geodynamic, and erosional processes that control the thermal history of the Himalayan orogenic interior since the onset of collision in early Paleogene time. In this paper, we present new (U-Th)/He thermochronometric cooling age data from 18 detrital zircons (ZHe) and 19 detrital apatites (AHe) of the early Eocene–early Miocene (ca. 50–23 Ma) continental facies of the Indus Group along the India-Asia collision zone in Ladakh, northwest (NW) India. This along-strike regional-scale low-temperature thermochronometric data set from the Indus basin is the first report of ZHe and AHe cooling ages from western and eastern Ladakh. Thermal modeling of our ZHe and AHe cooling ages indicates a postdepositional Neogene cooling signal in the Indus Group. Cooling initiated at ca. 21–19 Ma, was operational along the ~300 km strike of the collision zone in NW India by ca. 11 Ma, and continued until ca. 3 Ma. The Miocene cooling signal, also present along the India-Asia collision zone in south Tibet, is a continental-scale cooling event likely linked to increased erosional efficiency by the Indus and Yarlung Rivers across an elevated region resulting from the subduction dynamics of the underthrusting Indian plate.
in eastern Ladakh also record postdepositional Miocene cooling in the Indus Group [6, 13]. However, Tripathy-Lang et al. [10] obtained ca. 52–28 Ma predepositional unreset ZHe ages from the late Oligocene Basgo Formation of the Indus Group in central Ladakh. This finding raises a possibility that the cooling ages from the Indus basin may vary along-strike. Along the India-Asia collision zone in NW India (Figure 1), cooling ages from the Ladakh batholith vary along-strike: the western parts of the Ladakh batholith reveal ca. 46–41 Ma ZFT cooling ages [14], but ZHe, AFT, and AHe ages from the central part of the Ladakh batholith show discontinuous Eocene–Miocene cooling pulses from ca. 35 to 12 Ma [15, 16].

Low-temperature thermochronometric analyses using ZHe, AFT, and AHe age data reveal a Miocene cooling signal at multiple locations along the India-Asia collision zone in south Tibet between Kailas and Zedang (Figure 1(b)). Miocene cooling between 21 and 6 Ma is well documented along the strike of the collision zone in south Tibet across multiple tectonic units, including the Gangdese batholith, Kailas basin, Xigaze forearc, and the northern Tethyan Himalaya [17–26]. Miocene cooling in south Tibet occurred as the subducting Indian plate switched from rollback to underthrusting in early Miocene time and is attributed to regional dynamic uplift linked to Greater Indian slab breakoff [24, 27], movement on the GCT and/or Yarlung River incision (e.g., [21, 23]).

Therefore, to determine whether the Miocene cooling signal is also present along the strike of the collision zone in NW India, we analyzed and thermally modeled a total of 37 new single-grain detrital ZHe and AHe ages from four samples of the Indus Group in western, central, and eastern Ladakh (Figures 2–4). In addition, we compile 244 published AHe, AFT, and ZHe mean cooling ages (Figures 5(a)–5(c)).
and compare these to regional geologic observations to determine if Miocene cooling was a continental-scale thermal event along >1600 km of the India-Asia collision zone between Lehdo, western Ladakh, and Zedang, southeast Tibet (Figure 1(b)).

### 2. Geologic Background

A critical element of the India-Asia collision zone in Ladakh, NW India, is the Indus Group that is a part of the Indus basin sedimentary rocks. These sedimentary rocks unconformably overlie the Cretaceous–Paleogene Ladakh batholith to the north and truncate against a complex assemblage of Indus suture units to the south (Figure 1(a)). The Indus suture units consist of the Cretaceous Dras-Nidar Complexes, containing ophiolitic mélangé, volcanic and volcano-sedimentary rocks [28–30], and the Triassic–Cretaceous Lamayuru Complex, containing the continental slope facies of the northern Indian passive margin [31]. South of the Indus suture lies the Tethyan Himalaya, which is composed of a metasedimentary Neoproterozoic–early Cenozoic continental shelf succession of the northern Indian passive margin [32]. A klippe preserves the Mesozoic Spongtag oceanic arc within the Tethyan Himalaya (Figure 1; [33]). In the southeast, the subducted-extruded margin of the Indian continental plate is preserved in the Tso Morari Complex, which is an ultrahigh-pressure metamorphic unit [34]. A series of strike-parallel north-verging thrusts, linked to the Great Counter thrust (GCT) system, deformed the collision zone between ca. 23 and 20 Ma causing inversion of the Indus basin (Figure 1; [6, 9]).

The Indus basin sedimentary rocks consist of two rocks groups: the marine Tar Group, deposited in a forearc setting between Late Cretaceous and early Eocene time (Figure 2(a))

### Figure 2: Regional Indus Group stratigraphy in Ladakh, NW India (modified after Sinclair and Jaffey [50]), showing sampled formations, and ZHe-AHe thermochronology results reported in this study. Biostratigraphic ages of the marine strata and the Basgo Formation are from Green et al. [35] and Bajpai et al. [4]. U-Pb detrital zircon maximum depositional ages, calculated using YC2σ(3+) method, are from Bhattacharya et al. [5]. n is the number of zircon or apatite grains dated from a specific formation. Fm: formation.
and the continental Indus Group, deposited after the onset of India-Asia collision in an intermontane basin between early Eocene and early Miocene time (ca. 50–23 Ma; Figure 2(b); [4–8, 11]). Stratigraphically, the Indus Group consists of six regional-scale siliciclastic formations—the oldest Nurla Formation, the Choksti, Hemis, Nimu, and Basgo Formations, and the youngest Temesgam Formation (Figure 2). Deposition of continental sedimentary rocks also occurred locally in the Ladakh region in Pliocene time [7, 36].

Discrepancies exist concerning the timing of sedimentary rock exhumation among the few previous low-temperature detrital thermochronometric studies from the Indus basin in NW India that are limited mostly to central Ladakh. The cooling age data from the same rock group or formations using the same low-temperature thermochronometric techniques are significantly different. Tripathy-Lang et al. [10] determined ca. 52–28 Ma unreset pre depositional ZHe ages from the late Oligocene Basgo Formation in central Ladakh and suggest that the ages reflect cooling of Indian plate.
source regions. However, a recent study by Bhattacharya et al. [12], across four sections through the different Indus basin sedimentary rocks of the India-Asia collision zone (including the Basgo Formation) in central Ladakh, reveals postdepositional Miocene ZHe cooling ages from ca. 19 to 8 Ma and Miocene–Pliocene AHe cooling ages from ca. 7 to 4 Ma. These ZHe and AHe ages are interpreted to reflect Miocene exhumation along the Indus suture in response to Indus River erosion. Clift et al. [6] determined two AFT central ages of ca. 14–12 Ma from the early Eocene Nurla Formation in central Ladakh and associated them with postdepositional cooling following Indus basin inversion caused by the overthrusting GCT between ca. 23 and 20 Ma. In eastern Ladakh, Schlup et al. [13] report a single AFT central age of ca. 7 Ma from the Indus Group that is also interpreted to reflect postdepositional cooling. In addition, the central part of the neighboring Ladakh batholith yields ca. 26–18 Ma ZFT, ZHe, and AFT ages, which indicate rapid cooling in late Oligocene–early Miocene time along the collision zone [16]. Partially reset zircon fission-track (ZFT) ages from the Indus Group in central Ladakh suggest that peak basin temperatures exceeded the lower partial annealing zone temperatures of 185–200°C [12]. Basin temperature estimates from additional thermal proxies differ by tens of degrees. Vitritine reflectance and illite crystallinity techniques yield a maximum basin temperature of 155°C for the deepest Nurla Formation of the Indus Group in central Ladakh [37]. Illite crystallinity maximum basin temperature estimates of Clift et al. [6] from central Ladakh are between 175 and 200°C. Using the illite crystallinity index-temperature equation of Zhu et al. [38], the illite crystallinity analyses of Clift et al. [39] from eastern Ladakh yields a maximum basin temperature estimate of ~239°C. Overall, there is limited evidence that suggests Miocene cooling along the India-Asia collision zone in NW India might have been regional in scale. Alternatively, it is possible that the Indus Group experienced different burial and exhumation histories along-strike as demonstrated by the Paleogene unreset ZHe cooling ages of Tripathy-Lang et al. [10].

Figure 4: Plot showing the Miocene–Pliocene ZHe-AHe cooling ages from the Indus Group determined in this study and the timing of key geological processes that were occurring in the India-Asia collision zone in NW India and south Tibet in Miocene time. Solid circles and solid squares indicate ZHe and AHe cooling ages of the corresponding formation. Shaded area (light grey) indicates Miocene time. References: (1) Clift et al. [6], (2) Zhang et al. [66], (3) Sinclair and Jaffey [50], (4) Clift et al. [57], (5) Replumaz et al. [67], (6) DeCelles et al. [27], (7) Laskowski et al. [18], (8) Dèzes et al. [61], and (9) Shen et al. [24].
Figure 5: (a) Map of the India-Asia collision zone showing AHe, AFT, and ZHe sample locations compiled from NW India and south Tibet including this study. Abbreviations: GCT: Great Counter thrust; STDS: South Tibetan Detachment System. (b) Longitudinal variation of AHe, AFT, and ZHe ages from the Indus Group in NW India and the Kailas-Xigaze basins in south Tibet along the India-Asia collision zone. (c) Longitudinal variation of AHe, AFT, and ZHe ages from the overall India-Asia collision zone (including Indus, Kailas and Xigaze basins, Ladakh-Gangdese batholiths, and Tethyan Himalaya with its leucogranite belt). Grey box in B and C indicates Miocene time. Insets in B and C show Kernel Density Estimation curves (pink) for compiled AHe, AFT, and ZHe ages on frequency histogram charts. Peak age (Ma) is displayed alongside (see Table S5 for sources of ages corresponding to GPS locations).
3. Sample Collection and Methods

The primary technique used is (U-Th)/He detrital ZHe and AHe thermochronology. Sampling the Indus Group for detrital ZHe and AHe dating is challenging. Multiple sections preserve the basal, middle, and top parts of the Indus Group in Ladakh and a single section across the composite Indus Group stratigraphy is not accessible. We collected 4 detrital samples (DZ-prefix), one from each of the four key regional-scale Indus Group units: sample DZT2MH78 from the Nurla Formation (base) in eastern Ladakh, sample DZAR2HMS77 from the Hemis Formation (middle) in central Ladakh, and samples DZTSGK02 and DZASKR16 from the Basgo (middle) and Temesgam (top) Formations, respectively, in western Ladakh (Figure 2). DZT2MH78 and DZASKR16 are fine-grained sandstone samples, while DZAR2HMS77 and DZTSGK02 are very coarse-grained sandstone samples that were interbedded with conglomerates.

3.1. (U-Th)/He Thermochronology. Zircons and apatites were separated using standard techniques of crushing, grinding, magnetic, and heavy liquid separation. Four to six dateable quality sharp-faced euhedral zircon and apatite crystals with few or no inclusions were hand-picked from each sample and analyzed at the Arizona Radiogenic Helium Dating Laboratory following the procedures of Reiners et al. [40]. A total of 18 zircons and 19 apatites were dated. Selected zircon and apatite grains were packed in Nb tubes and subjected to standard procedures of He extraction analysis using a quadrupole mass spectrometer. Th and U contents were measured using ICP-MS following the methods in Reiners et al. [40]. Subsequently, raw ages were calculated by incorporating known analytical concentrations of U, Th, and He in the combined decay-diffusion equation and corrected by applying the alpha-ejection protocols [41, 42]. If resulting ZHe or AHe cooling ages are older than the depositional age of the formation, then these cooling ages are thermally unreset after deposition and reflect predepositional cooling in the source regions. In contrast, ZHe and AHe cooling ages that are younger than the depositional age of the formation suggest burial temperatures in the basin exceeded 140–200°C and 40–90°C, respectively, and indicate that these ages are therefore thermally reset after deposition. However, the (U-Th)/He system is sensitive to radiation damage, crystal anisotropy, grain size, annealing mechanism, and uncertainties in burial conditions, thereby reducing the precision associated with inferred thermal histories from ZHe or AHe cooling ages [43, 44]. We analyzed the variation of ZHe and AHe grain ages with effective uranium and grain size to identify if radiation damage and/or crystal size have a control on the observed distribution of cooling ages.

3.2. Thermal Modeling. Using the thermal modeling program HeFTy v.1.9.1 [45], the ZHe and AHe ages from each sample were inverse modeled to determine the time-temperature (t-T) history of the corresponding sample. For a given input data set that include uncorrected cooling ages, U-Th-Sm concentrations, grain size, and a given diffusion model (e.g., [44, 46]), the inverse algorithm extracted a family of t-T paths that a sample could have possibly experienced under the user assigned t-T constraints. Based on the calculated statistical fit between the measured and predicted cooling ages, the t-T pathways generated for a particular sample can be grouped into two sets—the acceptable-fit paths that have a Kolmogorov-Smirnov probability ≥0.05 and the good-fit paths, which have a Kolmogorov-Smirnov probability ≥0.5. The t-T path with the highest goodness of fit is the best-fit path. The best-fit path is sensitive to the number of paths tested but reasonably approximates a meaningful thermal history experienced by a particular sample when a sufficiently high number of paths are tested.

Applying the inverse algorithm of the thermal modeling program HeFTy, we modeled the t-T histories of the four samples using the ZHe and AHe cooling age data and published regional geologic t-T constraints. We used depositional age constraints as discussed in “Constraining Depositional Ages” and considered a surface depositional temperature of 0–25°C for all samples. No evidence of cooling exists in the Indus Group before 23 Ma, when overthrusting by the GCT inverted the Indus basin [6]. Recent ZFT age analyses from the Indus Group suggest that basin temperatures exceeded the lower partial annealing zone temperatures of 185–200°C but stayed below higher partial annealing zone temperatures of 280–300°C [5]. An illite crystallinity estimate by Clift et al. [39] also indicates a basin temperature of ~239°C from the Indus Group. Thus, based on regional geologic constraints, we allowed each model to explore the t-T space younger than 23 Ma and colder than 240°C. Considering a 240°C temperature limit also permits the model to extract t-T histories from temperatures greater than the closure temperature range of the warmest thermochronometer in the study, i.e., ZHe system, thereby relaxing the t-T space being investigated. We used mean ZHe and AHe ages in modeling the t-T histories of the samples using the diffusion models of Guenther et al. [44] and Farley [46]. Using the same t-T constraints, we also modeled the single-grain ZHe and AHe ages from each sample, as well as all the single-grain ZHe ages individually, to test if the resultant models exhibit thermal histories that are significantly different from those obtained by using mean ZHe and AHe ages. We ran all models for >800,000 paths and until at least 100 good-fit paths were obtained and considered the best-fit path that resulted at the end of each model run to reasonably approximate a robust t-T history under the applied geologic constraints.

3.3. Constraining Depositional Ages. Limited fossil data from the Indus Group formations make it difficult to precisely estimate their true depositional ages, which are critical initial constraints in thermal modeling. We use the depositional ages of the studied Indus Group formations from published biostratigraphic ages [7, 35], detrital zircon U-Pb maximum depositional ages [5, 7, 11] and ages of tectonic events that paused deposition [6]. The depositional age of the Nurla Formation, from which sample DZT2MH78 is collected, is ca. 50–41 Ma. The lower age limit of 50 Ma is constrained by the ca. 50 Ma U-Pb detrital zircon maximum depositional age of the Nurla Formation as well as the fossils of same
age from the marine sequence immediately underlying the Nurla Formation [5, 35]. The upper age limit of 41 Ma is imposed from the ca. 46–41 Ma U-Pb detrital zircon maximum depositional age of the overlying Choksti Formation [7, 11]. The 41 Ma upper age limit for the Nurla Formation also coincides with an unconformity confirming termination of Nurla deposition [47]. The depositional age of the Hemis Formation, from which sample DZAR2HMS77 is collected, is constrained to ca. 37–28 Ma. The lower age limit of 37 Ma comes from the detrital zircon U-Pb maximum depositional age of the Hemis Formation [5]. The upper age limit of 28 Ma comes from the late Oligocene biostratigraphic age of the Basgo Formation [4], which is younger than the Hemis Formation. It is possible that Hemis deposition terminated earlier; however, the lack of fossils in the Hemis Formation or in the overlying Nimu Formation (ca. 32 Ma detrital zircon U-Pb maximum depositional age of ca. 27 Ma [5], is constrained to ca. 28–26 Ma. Stratigraphically, the Basgo Formation is ~10200 m thick [48] and is conformably overlain by the Temesgam Formation that has a detrital zircon U-Pb maximum depositional age of ca. 26 Ma [5]. The depositional age of the Temesgam Formation is therefore constrained to ca. 26–23 Ma, with the upper age limit of ca. 23 Ma from the estimated age of counterthrusting by the GCT that ended regional Indus Group deposition [6].

4. Results and Interpretation

The ZHe and AHe cooling ages from the Indus Group are ca. <20 Ma (Figure 2; Tables S1 and S2). The ZHe ages are between 19.65 ± 0.25 and 9.37 ± 0.12 Ma with most of the ages between ca. 16 and 10 Ma. The AHe ages are between 9.57 ± 0.11 and 3.23 ± 0.11 Ma with majority of the ages between ca. 7 and 3 Ma. Sample DZT2MH78 from the stratigraphically deepest Nurla Formation in eastern Ladakh displays ZHe and AHe ages from 18.03 ± 0.22 to 11.00 ± 0.19 Ma and from 4.42 ± 0.07 to 3.23 ± 0.11 Ma, respectively. Corresponding ZHe and AHe age ranges from sample DZAR2HMS77 of the Hemis Formation in central Ladakh are 19.45 ± 0.24–13.55 ± 0.17 Ma and 9.57 ± 0.11–5.57 ± 0.10 Ma, respectively. In western Ladakh, sample DTZTSG02 from the Basgo Formation exhibits ZHe and AHe ages from 19.65 ± 0.25 to 15.67 ± 0.20 Ma and from 7.13 ± 0.16 to 3.88 ± 0.09 Ma, respectively. Corresponding ZHe and AHe ages from sample DZASKR16 of the stratigraphically shallowest Temesgam Formation in western Ladakh range from 14.98 ± 0.19 to 9.37 ± 0.12 Ma and from 6.81 ± 0.10 to 4.59 ± 0.10 Ma, respectively.

The ca. 20–3 Ma ZHe and AHe cooling ages from the Nurla, Hemis, Basgo, and Temesgam Formations are all younger than their corresponding ca. 50–41 Ma, ca. 37–28 Ma, ca. 28–26 Ma, and ca. 26–23 Ma depositional ages, and the overall ca. 50–23 Ma depositional age of the Indus Group [5]. This implies thermal resetting of both ZHe and AHe systems and requires burial of the Indus Group to at least >140-200°C after deposition [44].

Within individual samples, correlations between grain size and ZHe or AHe ages are either very weak or nonexistent. However, considering the 18 ZHe ages together, a moderately positive correlation between grain size and age accounts for the ZHe age dispersion in the samples (Figure S3). Likewise, the 19 AHe ages together show a weak positive correlation with grain size, which possibly explains the intrasample AHe age dispersion (Figure S3). No meaningful correlations exist between effective uranium and grain ages in individual samples or collectively for either thermochronometric system, suggesting radiation damage is not the primary control on observed intrasample ZHe and AHe age variability. Tables S1 and S2 and Figure S3 contain all analytical data, and cooling age versus grain size and effective uranium plots.

5. Thermal Modeling

Modeling individual ZHe and AHe ages for a sample, i.e., 8-10 single-grain ages together in HeFTy resulted in no acceptable or good-fit paths. That HeFTy fails to yield meaningful thermal histories by simultaneously solving all input data from multiple grain ages of the same sample is known (e.g., [17]). The t-T modeling results using mean ZHe and AHe ages are in Figures 3(a)–3(d), while the t-T modeling results from the 18 individual ZHe ages are in Figure S4. Because these two sets of t-T modeling results do not differ significantly, we continue the discussion with reference to the t-T models using mean ages from here on.

Thermal modeling of ZHe and AHe ages confirms thermal resetting of both thermochronometric systems and exhibits best-fit t-T paths that suggest initiation of postdepositional cooling between ca. 21 and 11 Ma in all samples (Figures 3(a)–3(d)). Cooling begins in the Hemis and Basgo Formations at ca. 21–19 Ma, while cooling begins later in the Nurla and Temesgam Formations at ca. 15–11 Ma. As shown by several acceptable and good-fit t-T paths in each model, cooling may have initiated earlier than the time indicated by the best-fit t-T path in the corresponding model. In this study, we consider the timing of beginning of cooling indicated by the best-fit t-T path as the minimum time by which cooling began in each model. The best-fit t-T paths in Nurla and Hemis Formations indicate cooling beginning from peak basin temperatures of ~225°C (Figures 3(a) and 3(b)). The best-fit t-T paths in the shallower Basgo and Temesgam Formations indicate peak basin temperatures of ~150 and 180°C, respectively (Figures 3(c) and 3(d)). The relatively low peak basin temperatures in the Basgo and Temesgam Formations are likely a consequence of modeling. Because the younger Temesgam Formation was at a higher elevation than the older Basgo Formation, the Temesgam Formation is expected to show cooler maximum basin temperatures. Nevertheless, modeling of the individual ZHe ages from the Basgo and Temesgam Formations indicate peak basin temperatures exceeding 180–200°C (Figure S4).

The acceptable and good-fit t-T paths appear to pass through either higher temperatures (>180–200°C) early or through lower temperatures (<180–200°C) later resulting in
two distinct clusters of $t$-$T$ solutions in each model. From a strict modeling point of view, these two distinct clusters of solutions indicate that either can satisfactorily explain the two input data points, i.e., a mean ZHe age and a mean AHe age, and constrain the timing of initiation of cooling at any time between ca. 23 and 5 Ma. The older higher temperature cluster predicts the mean ZHe age, while the younger lower temperature cluster predicts the mean AHe age. The two clusters are essentially part of a single cooling event that initiated from temperatures >180–200°C, which can be independently verified from regional paleo-temperature estimates using illite crystallinity and ZFT analyses (e.g.,[12, 39]). The clusters are likely a consequence of averaging single-grain ZHe or AHe ages, and this problem may be avoided if more ZHe and AHe individual grain ages from a single sample are available to analyze the distribution of ages and predict each age using the method of Fox et al. [49]. However, acquiring good-quality dateable zircon or apatite grains was challenging, and given the limited ZHe and/or AHe ages in our individual samples, we use available data from regional thermal proxies and thermochronometric studies to validate the results of our $t$-$T$ modeling.

### 6. Discussion

The new ZHe and AHe cooling ages together with the thermal modeling reveal a postdepositional Miocene–Pliocene cooling phase initiated in the Indus Group by ca. 21–19 Ma and by 11 Ma was in operation along the ~300 km strike of the India-Asia collision zone from western to eastern Ladakh in NW India. The $t$-$T$ models also confirm that maximum burial temperatures in the Indus Group were close to or exceeded the upper limit of the ZHe partial retention temperatures, i.e., 180–200°C. This temperature estimate is consistent with the maximum burial temperature of ~239°C determined by Clift et al. [39] using illite crystallinity. Our study expands the limited regional low-temperature thermochronometric data set and shows that the postdepositional Miocene cooling signal in the Indus basin is present from western to eastern Ladakh. This implies that the basin did not undergo differential burial and exhumation histories along the length of the India-Asia collision zone in NW India. The new ca. 20–3 Ma ZHe and AHe ages from western to eastern Ladakh overlap with the published ca. 19–8 Ma ZHe, 14–12 Ma AFT, and 7–3 Ma AHe ages from the Indus Group that infer a postdepositional Miocene–Pliocene cooling signal in central Ladakh [6, 12]. Our thermal models and the ca. 10–3 Ma AHe ages also indicate that the cooling phase continued into Pliocene time at least until ca. 3 Ma. It is difficult to explain why the ZHe cooling ages from two different locations of the Basgo Formation in this study and in the work by Bhattacharya et al. [12] differ from those determined by Tripathy-Lang et al. [10] from the same formation. Nevertheless, our ZHe cooling ages from the Basgo Formation are compatible with the overall trend of the ZHe cooling ages determined from the Indus Group.

Across Ladakh, the late Oligocene–early Miocene Temesgam Formation is the youngest Indus Group stratigraphy on a regional scale and has postdepositional ZHe and AHe cooling ages like the deeper and older stratigraphic units. Localized Pliocene deposition [7], restricted to the Zanskar Gorge area of central Ladakh, is <1 km in stratigraphic thickness and cannot explain the postdepositional cooling ages in the Temesgam Formation. Therefore, sedimentation alone cannot account for postdepositional burial temperatures >180–200°C in the ~4.5 km thick Indus Group. We suggest that the Indus basin sedimentary rocks, which contains the Indus Group, were buried by the GCT hanging wall rocks between ca. 23 and 20 Ma [6, 12, 50]. Although thermochronometric data do not directly constrain surface uplift, the timing of Miocene cooling in the Indus Group coincides with key geodynamic and erosional processes operating in the India-Asia collision zone at that time that are linked to regional surface uplift and denudation (Figure 4). The Indus basin was affected by a combination of contemporaneous geodynamic (e.g., Indian slab dynamics), tectonic (e.g., slip along the GCT), and erosional processes (e.g., Indus River incision) in Miocene time in NW India. We suggest that, following Greater India slab rollback and breakoff, the return of northward underthrusting of the Indian plate in early Miocene time provided the mechanism for regional surface uplift of the overall India-Asia collision zone [24, 27, 51–53], including the GCT-overthrust Indus basin sedimentary rocks. This surface uplift was responsible for vertical elevation gains across the basin. The present-day elevations of >4–5 km of the Indus basin are therefore a direct consequence of the northward underthrusting of the Indian plate that uplifted the collision zone and crustal shortening along the GCT, although the relative contributions to elevation gains by these two processes in NW India are not yet determined. The Indus River flowed through the elevated region in Neogene time eroding the overthrust Indus basin rocks and producing the observed Miocene cooling signal. Considering cooling from a mean maximum burial temperature of 200°C and assuming a geothermal gradient of 20–30°C/km that is consistent with the regional Miocene geothermal gradients estimated from neighboring lithotectonic units [54–56], we infer a removal of ~7–10 km of rock from the Indus Group since the onset of Miocene cooling. Chemical weathering and alteration indices and sedimentation rates determined from Indus fan sediments show that erosion peaked in the Himalaya at ca. 15–10 Ma transporting rocks from the orogenic front to the Indus fan [57]. Our rock erosional estimate suggests that, during Miocene time, the orogenic hinterland contributed a significant amount of rock material to the Indus fan as well. The caveat here is that the Indus River might not have been the sole driver of regional erosion. The Asian monsoon, which intensified during Miocene time in the orogenic hinterland [57], may have played a role in erosion and subsequent cooling as well. However, more studies are required in the region to investigate the interplay of multiple drivers of erosion.

Compared to NW India, low-temperature thermochronometric studies in the India-Asia collision zone of south Tibet are abundant and unveil the regional thermal history. ZHe cooling ages from the late Oligocene–early Miocene Kailas Formation in south Tibet, which is contemporaneous with the upper formations of the Indus Group in NW India,
reveal basin exhumation between ca. 21 and 15 Ma [17]. AFT and AHe cooling ages indicate that the Cenozoic Liuqu Formation experienced cooling after ca. 12–10 Ma [22]. Thermal modeling of ZHe and AHe ages, and AFT age data show that the Cretaceous–early Eocene Xigaze forearc cooled in Miocene time from ca. 21–7 Ma [20, 23]. ZHe, AFT, and AHe thermochronology of the adjacent southern Gangdese batholith and the northern Tethyan Himalaya in south Tibet also depict cooling between ca. 20 and 6 Ma [18, 19, 21, 26, 58]. The Miocene cooling signal along the India-Asia collision zone in south Tibet is attributed to regional dynamic surface uplift that accompanied northward underthrusting of the Indian plate after Greater Indian slab rollback and breakoff [24, 27], Asian monsoon incision (e.g., [17]), movement on the GCT and/or Yarlung River (e.g., [17, 19, 20, 22, 23, 26, 58]).

Comparison of our thermochronometric data with a compilation of 244 published mean ZHe, AFT, and AHe cooling ages from NW India and south Tibet reveals that 75% of the ages are Miocene in age (Figures 5(a)–5(c)). The postdepositional ages of the sedimentary units along the India-Asia collision zone—the Indus, Kailas, Liuqu, and Xigaze basins—and the overall India-Asia collision zone exhibit similar cooling age distribution trends with a peak Miocene cooling age of ca. 17 Ma (Figures 5(b) and 5(c)). Although cooling ages may vary with grain size, radiation damage, presence of inclusions, and overall structural position, the ages in this compilation reflect erosional cooling through closure temperatures of ZHe (140–200°C, [44]) and AHe (40–90°C, Farley et al., 2000) corresponding to ~2–10 km depths, assuming 20–30°C/km geotherms. The combined data set of low-temperature thermochronometric ages suggests a continental-scale Miocene erosional cooling event along >1600 km of the India-Asia collision zone between Lehdo and Zedang (Figure 1(b)). Cooling may have initiated in Paleogene time in the nonsedimentary units (e.g., Ladakh-Gangdese batholiths, Tethyan Himalaya); however, Neogene time is when all tectonic units of the India-Asia collision zone were exhuming (Figures 5(b) and 5(c)). Given the timing of key geodynamic, tectonic, and erosional processes that coincide with the timing of Miocene cooling in NW India (Figure 4) and south Tibet, we hypothesize that a combination of these processes played a role in the observed Miocene cooling signal, i.e., the regional Miocene geology supports the thermochronology. In other words, the set of processes controlling cooling in Paleogene time (e.g., [55, 59, 60]) was different from those controlling cooling in Neogene time. The onset of Miocene cooling on a continental scale is coeval with the regional dynamic surface uplift and a return of northward underthrusting of the Indian plate following Greater India slab breakoff, and both these processes are well documented in south Tibet [24, 27]. Using ZFT-AFT thermochronology and topography modeling, Shen et al. [24] interpret that the Miocene cooling signal along the Yarlung suture in the India-Asia collision zone of south Tibet is a consequence of dynamic topography, by which a southerly migrating wave of topographic uplifts and subsidence reflect the surface response to sublithospheric mantle flow perturbations. If the dynamic topography hypothesis is applicable to NW India, this suggests that Miocene cooling occurred regionally across the Indus and Yarlung sutures along the India-Asia collision zone in response to dynamic topographic uplift associated with the lithospheric slab dynamics of the subducting Indian plate. The onset of Miocene cooling in NW India is also coeval with the ~22–19 Ma timing of normal faulting along the regional South Tibetan Detachment south of the study area (Figure 1(a); [61]) or immediately precedes regional orogen-parallel E-W extension, which initiated at ~16 Ma along the India-Asia collision zone in NW India and south Tibet [18, 62, 63]. However, low-temperature thermochronometric data from NW India are limited, and additional data are needed to test the dynamic topography hypothesis, or any potential cause-effect relationship between the onset of Miocene cooling and orogen-parallel extension or normal faulting along the South Tibetan Detachment. Nevertheless, the role of coeval lithospheric slab dynamics in NW India in relation to the regional surface uplift and observed Miocene cooling signal cannot be ruled out, although its exact nature remains poorly understood. Therefore, we suggest that the Miocene cooling along the India-Asia collision zone was exacerbated by downcutting erosion by the Indus and Yarlung Rivers through an elevated region produced by the northward underthrusting of the Indian slab after Greater India slab breakoff. Basins along the India-Asia collision zone were buried either through sedimentation and/or by the GCT system (e.g., [17, 19, 23]; this study) depending on local structural relationships. Interestingly, this pronounced Neogene cooling event along the India-Asia collision zone is contemporaneous with cooling to the south in Greater Himalayan rocks (e.g., [64]), and whether these two signals might possibly be related to large-scale evolution of the orogenic wedge remains a topic for future research.

7. Conclusions

Our new ZHe-AHe ages and thermal models from the Indus Group indicate a regional-scale Miocene–Pliocene cooling phase along the ~300 km strike of the India-Asia collision zone in NW India. Cooling initiated in the Indus Group in Miocene time at ca. 21–19 Ma following burial of the Indus basin by the GCT to >180–200°C and continued at least until ca. 3 Ma. The Miocene cooling signal in NW India is part of a continental-scale Miocene cooling signal observed along >1600 km of the India-Asia collision zone from Lehdo, NW India, to Zedang, south Tibet. Subduction dynamics, specifically a return to northward underthrusting of the Indian plate, coupled with erosion by the Indus and Yarlung Rivers in NW India and south Tibet, likely drove Neogene cooling across the India-Asia collision zone.

Data Availability

(U-Th)/He thermochronometric data presented in this paper are available in the supplementary file.
Conflicts of Interest

The authors do not report any conflicts of interest regarding the contents of this paper.

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Supplementary Materials

Table S1: (U-Th)/He detrital zircon (ZHe) age data table. Table S2: (U-Th)/He detrital apatite (AHe) age data table. Figure S3: plot of (U-Th)/He ZHe-AHe cooling ages vs effective uranium and grain size. Figure S5: time-temperature modeling of individual (U-Th)/He zircon ages using HEFTy. Table S5: compilation of published AHe, AFT, and ZHe ages from NW India and south Tibet. (Supplementary Materials)

References

[1] P. Kapp and P. G. DeCelles, “Mesozoic–Cenozoic geological evolution of the Himalayan-Tibetan orogen and working tectonic hypotheses,” American Journal of Science, vol. 319, no. 3, pp. 159–254, 2019.
[2] Y. Najman, “The detrital record of orogenesis: a review of approaches and techniques used in the Himalayan sedimentary basins,” Earth-Science Reviews, vol. 74, no. 1, pp. 1–72, 2005.
[3] M. P. Searle, “Timing of subduction initiation, arc formation, ophiolite obduction and India–Asia collision in the Himalaya,” Geological Society, London, Special Publications, vol. 483, no. 1, pp. 19–37, 2019.
[4] S. Bajpai, R. C. Whatley, G. V. R. Prasad, and J. E. Whittaker, “An Oligocene non-marine ostracod fauna from the Basgo Formation (Ladakh Molasse), NW Himalaya, India,” Journal of Micropalaeontology, vol. 23, no. 1, pp. 3–9, 2004.
[5] G. Bhattacharya, D. M. Robinson, and M. M. Wielicki, “Detrital zircon provenance of the Indus Group, Ladakh, NW India: Implications for the timing of the India-Asia collision and other syn-orogenic processes,” GSA Bulletin, vol. 133, no. 5-6, pp. 1007–1020, 2021.
[6] P. D. Clift, A. Carter, M. Krol, and E. Kirby, “Constraints on India-Eurasia collision in the Arabian Sea region taken from the Indus Group, Ladakh Himalaya, India,” Geological Society, London, Special Publications, vol. 195, no. 1, pp. 97–116, 2002.
[7] A. L. Henderson, Y. Najman, R. Parrish et al., “Geology of the Cenozoic Indus Basin sedimentary rocks: Paleoenvironmental interpretation of sedimentation from the western Himalaya during the early phases of India-Eurasia collision,” Tectonics, vol. 29, no. 6, 2010.
[8] A. L. Henderson, Y. Najman, R. Parrish, D. F. Mark, and G. L. Foster, “Constraints to the timing of India-Eurasia collision; a re-evaluation of evidence from the Indus Basin sedimentary rocks of the Indus-Tsangpo Suture Zone, Ladakh, India,” Earth-Science Reviews, vol. 106, no. 3-4, pp. 265–292, 2011.
[9] M. P. Searle, K. T. Pickering, and D. J. W. Cooper, “Restoration and evolution of the intermontane Indus molasse basin, Ladakh Himalaya, India,” Tectonophysics, vol. 174, no. 3-4, pp. 301–314, 1990.
[10] A. Tripathy-Lang, K. V. Hodges, M. C. van Soest, and T. Ahmad, “Evidence of pre-Oligocene emergence of the Indian passive margin and the timing of collision initiation between India and Eurasia,” Lithosphere, vol. 5, no. 5, pp. 501–506, 2013.
[11] F.-Y. Wu, P. D. Clift, and J.-H. Yang, “Zircon Hf isotopic constraints on the sources of the Indus Molasse, Ladakh Himalaya, India,” Tectonics, vol. 26, no. 2, 2007.
[12] G. Bhattacharya, D. M. Robinson, D. A. Orme, Y. Najman, and A. Carter, “Low-temperature thermochronology of the Indus Basin in Central Ladakh, Northwest India: implications of Miocene-Pliocene cooling in the India-Asia collision zone,” Tectonics, vol. 39, no. 10, article e2020TC006333, 2020.
[13] M. Schlup, A. Carter, M. Cosca, and A. Steck, “Exhumation history of eastern Ladakh revealed by 40Ar/39Ar and fission-track ages: the Indus River–Ts’o Morari transect, NW Himalaya,” Journal of the Geological Society, vol. 160, no. 3, pp. 385–399, 2003.
[14] R. B. Sorkhabi, A. K. Jain, S. Nishimura et al., “New age constraints on the cooling and unroofing history of the Trans-Himalayan Ladakh Batholith (Kargil area), N.W. India,” Proceedings of the Indian Academy of Sciences - Earth and Planetary Sciences, vol. 103, no. 83–97, 1994.
[15] L. A. Kirstein, J. P. T. Foeken, P. van der Beek, F. M. Stuart, and R. J. Phillips, “Cenozoic unroofing history of the Ladakh Batholith, western Himalaya, constrained by thermochronology and numerical modelling,” Journal of the Geological Society, vol. 166, no. 4, pp. 667–678, 2009.
[16] L. A. Kirstein, H. Sinclair, F. M. Stuart, and K. Dobson, “Rapid early Miocene exhumation of the Ladakh batholith, western Himalaya,” Geology, vol. 34, no. 12, pp. 1049–1052, 2006.
[17] B. Carrapa, D. A. Orme, P. G. DeCelles, P. Kapp, M. A. Cosca, and R. Waldrip, “Miocene burial and exhumation of the India-Asia collision zone in southern Tibet: response to slab dynamics and erosion,” Geology, vol. 42, no. 5, pp. 443–446, 2014.
[18] A. K. Laskowski, P. Kapp, L. Ding, C. Campbell, and X. Liu, “Tectonic evolution of the Yarlung suture zone, Lopu Range region, southern Tibet,” Tectonics, vol. 36, no. 1, pp. 108–136, 2017.
[19] G. Li, B. Kohn, M. Sandiford, Z. Ma, and Z. Xu, “Post-collisional exhumation of the Indus-Yarlung suture zone and Northern Tethyan Himalaya, Saga, SW Tibet,” Gondwana Research, vol. 64, pp. 1–10, 2018.
[20] G. Li, B. Kohn, M. Sandiford, and Z. Xu, “India-Asia convergence: insights from burial and exhumation of the Xigaze fore-arc basin, South Tibet,” Journal of Geophysical Research: Solid Earth, vol. 122, no. 5, pp. 3430–3449, 2017.
[21] G. Li, B. Kohn, M. Sandiford, Z. Xu, Y. Tian, and C. Seiler, “Synorogenic morphotectonic evolution of the Gangdese batholith, South Tibet: insights from low-temperature thermochronology,” Geochemistry, Geophysics, Geosystems, vol. 17, no. 1, pp. 101–112, 2016.
[22] G. Li, B. Kohn, M. Sandiford, Z. Xu, and L. Wei, “Constraining the age of Liuqu Conglomerate, southern Tibet: Implications
for evolution of the India–Asia collision zone,” *Earth and Planetary Science Letters*, vol. 426, pp. 259–266, 2015.

[23] D. A. Orme, “Burial and exhumation history of the Xigaze forearc basin, Yarlung suture zone, Tibet,” *Geoscience Frontiers*, vol. 10, no. 3, pp. 895–908, 2019.

[24] T. Shen, G. Wang, A. Replumaz et al., “Miocene subsidence and surface uplift of southernmost Tibet induced by Indian subduction dynamics,” *Geochemistry, Geophysics, Geosystems*, vol. 21, no. 10, article e2020GC009078, 2020.

[25] T. Shen, G. Wang, P. H. Leloup et al., “Controls on Cenozoic exhumation of the Tethyan Himalaya from fission-track thermochronology and detrital zircon U-Pb geochronology in the Gyirong basin area, southern Tibet,” *Tectonics*, vol. 35, no. 7, pp. 1713–1734, 2016.

[26] M. M. Tremblay, M. Fox, J. L. Schmidt et al., “Erosion in southern Tibet shut down at ~10 Ma due to enhanced rock uplift within the Himalaya,” *Proceedings of the National Academy of Sciences*, vol. 112, no. 39, pp. 12030–12035, 2015.

[27] P. G. DeCelles, P. Kapp, J. Quade, and G. E. Gehrels, “Miocene–Miocene Kailas basin, southwestern Tibet: record of postcollisional upper-plate extension in the Indus-Yarlung suture zone,” *GSA Bulletin*, vol. 123, no. 7-8, pp. 1337–1362, 2011.

[28] T. Ahmad, T. Tanaka, H. K. Sachan, Y. Asahara, R. Islam, and P. P. Khanna, “Geochemical and isotopic constraints on the age and origin of the Nidar Ophiolitic Complex, Ladakh, India: Implications for the Neo-Tethyan subduction along the Indus suture zone,” *Tectonophysics*, vol. 451, no. 1-4, pp. 206–224, 2008.

[29] P. D. Clift, P. J. Degnan, R. Hannigan, and J. Blusztajn, “Sedimentary and geochemical evolution of the Dras forearc basin, Indus suture, Ladakh Himalaya, India,” *Geological Society of America Bulletin*, vol. 112, no. 3, pp. 450–466, 2000.

[30] S. Das, A. R. Basu, B. K. Mukherjee et al., “Origin of Indus ophiolite-hosted ophicarbonate veins: Isotopic evidence of mixing between seawater and continental crust-derived fluid during Neo-Tethys closure,” *Chemical Geology*, vol. 551, article 119772, 2020.

[31] A. Robertson and I. Sharp, “Mesozoic deep-water slope/-rise sedimentation and volcanism along the North-Indian passive margin: evidence from the Karamba Complex, Indus suture zone (Western Ladakh Himalaya),” *Journal of Asian Earth Sciences*, vol. 16, no. 2, pp. 195–215, 1998.

[32] E. Garzanti, A. Baud, and G. Mascle, “Sedimentary record of the northward flight of India and its collisions with Eurasia (Ladakh Himalaya, India),” *Geodinamica Acta*, vol. 1, no. 4-5, pp. 297–312, 1987.

[33] E. J. Catlos, E. C. Pease, N. Dygert et al., “Nature, age and emplacement of the Spongtag ophiolite, Ladakh, NW India,” *Journal of the Geological Society*, vol. 176, no. 2, pp. 284–305, 2019.

[34] D. Dutta and S. Mukherjee, “Extrusion kinematics of UHP terrane in a collisional orogen: EBSB and microstructure-based approach from the Tso Morari Crystallines (Ladakh Himalaya),” *Tectonophysics*, vol. 800, article 228641, 2021.

[35] O. R. Green, M. P. Searle, R. I. Corfield, and R. M. Corfield, “Cretaceous–Tertiary carbonate platform evolution and the age of the India–Asia collision along the Ladakh Himalaya (Northwest India),” *The Journal of Geology*, vol. 116, no. 4, pp. 331–353, 2008.

[36] N. Mathur, “The Indus Formation of the Ladakh Himalaya: its biozonation, correlation and faunal provincialism,” *Geology of Indus Suture Zone of Ladakh*, pp. 127–144, 1983.

[37] T. Van Haver, *Etude stratigraphique, sédimentologique et structurale d’un bassin d’avant arc: exemple du bassin de l’Indus, Ladakh, Himalaya*, Université Scientifique et Médicale de Grenoble, 1984, https://tel.archives-ouvertes.fr/tel-00641418.

[38] C. Zhu, S. Rao, and S. Hu, “Application of illite crystallinity for paleo-temperature reconstruction: a case study in the western Sichuan basin, SW China,” *Carpathian Journal of Earth and Environmental Sciences*, vol. 11, no. 2, pp. 599–608, 2016.

[39] P. D. Clift, M. Schlup, A. Carter, and A. Steck, “Discussion of exhumation history of eastern Ladakh revealed by40Ar/39Ar and fission track ages: the Indus River–Tso Morari transect, NW Himalaya,” *Journal of the Geological Society*, vol. 161, no. 5, pp. 893–894, 2004.

[40] P. W. Reiners, T. L. Spell, S. Nicoleus, and K. A. Zanetti, “Zircon (U-Th)/He thermochronometry: He diffusion and comparisons with 40Ar/39Ar dating,” *Geochemica et Cosmochimica Acta*, vol. 68, no. 8, pp. 1857–1887, 2004.

[41] K. A. Farley, “(U-Th)/he dating: techniques, calibrations, and applications,” *Reviews in Mineralogy and Geochemistry*, vol. 47, no. 1, pp. 819–844, 2002.

[42] J. K. Hourigan, P. W. Reiners, and M. T. Brandon, “U-Th zonation-dependent alpha-ejection in (U-Th)/He chronometry,” *Geochemica et Cosmochimica Acta*, vol. 69, no. 13, pp. 3349–3365, 2005.

[43] M. Fox and D. L. Shuster, “The influence of burial heating on the (U-Th)/He system in apatite: Grand Canyon case study,” *Earth and Planetary Science Letters*, vol. 397, pp. 174–183, 2014.

[44] W. R. Guenthner, P. W. Reiners, R. A. Ketcham, L. Nasdala, and G. Giester, “Helium diffusion in natural zircon: radiation damage, anisotropy, and the interpretation of zircon (U-Th)/He thermochronology,” *American Journal of Science*, vol. 313, no. 3, pp. 145–198, 2013.

[45] R. A. Ketcham, “Forward and inverse modeling of low-temperature thermochronometry data,” *Reviews in Mineralogy and Geochemistry*, vol. 58, no. 1, pp. 275–314, 2005.

[46] K. A. Farley, “Helium diffusion from apatite: general behavior as illustrated by Durango fluorapatite,” *Journal of Geophysical Research: Solid Earth*, vol. 105, no. B2, pp. 2903–2914, 2000.

[47] M. R. St-Onge, N. Rayner, and M. P. Searle, “Zircon age determinations for the Ladakh batholith at Chumathang (Northwest India): implications for the age of the India-Asia collision in the Ladakh Himalaya,” *Tectonophysics*, vol. 495, no. 3-4, pp. 171–183, 2010.

[48] E. Garzanti and T. Van Haver, “The indus clastics: forearc basin sedimentation in the Ladakh Himalaya (India),” *Sedimentary Geology*, vol. 59, no. 3-4, pp. 237–249, 1988.

[49] M. Fox, J.-G. Dai, and A. Carter, “Badly behaved detrital (U-Th)/He ages: problems with he diffusion models or geological models?,” *Geochemistry, Geophysics, Geosystems*, vol. 20, pp. 2418–2432, 2019.

[50] H. D. Sinclair and N. Jaffey, “Sedimentology of the Indus Group, Ladakh, northern India: implications for the timing of initiation of the palaeo-Indus River,” *Journal of the Geological Society*, vol. 158, no. 1, pp. 151–162, 2001.
[51] P. G. DeCelles, I. S. Castañeda, B. Carrapa et al., "Oligocene-Miocene Great Lakes in the India-Asia Collision Zone," *Basin Research*, vol. 30, no. S1, pp. 228–247, 2018.

[52] A. A. G. Webb, H. Guo, P. D. Clift et al., "The Himalaya in 3D: slab dynamics controlled mountain building and monsoon intensification," *Lithosphere*, vol. 9, no. 4, article L636.1, 2017.

[53] Q. Xu, L. Ding, R. A. Spicer, X. Liu, S. Li, and H. Wang, "Stable isotopes reveal southward growth of the Himalayan-Tibetan Plateau since the Paleocene," *Gondwana Research*, vol. 54, pp. 50–61, 2018.

[54] J.-L. Epard and A. Steck, "Structural development of the Tso Morari ultra-high pressure nappe of the Ladakh Himalaya," *Tectonophysics*, vol. 451, no. 1-4, pp. 242–264, 2008.

[55] R. Kumar, A. K. Jain, N. Lal, and S. Singh, "Early–Middle Eocene exhumation of the Trans-Himalayan Ladakh Batholith, and the India–Asia convergence," *Current Science*, vol. 113, no. 6, pp. 1090–1098, 2017.

[56] M. Schlup, A. Steck, A. Carter, M. Cosca, J.-L. Epard, and J. Hunziker, "Exhumation history of the NW Indian Himalaya revealed by fission track and $^{40}$Ar/$^{39}$Ar ages," *Journal of Asian Earth Sciences*, vol. 40, no. 1, pp. 334–350, 2011.

[57] P. D. Clift, K. V. Hodges, D. Heslop, R. Hannigan, H. Van Long, and G. Calves, "Correlation of Himalayan exhumation rates and Asian monsoon intensity," *Nature Geoscience*, vol. 1, no. 12, pp. 875–880, 2008.

[58] Y.-K. Ge, J.-G. Dai, C.-S. Wang, Y.-L. Li, G.-Q. Xu, and M. Danisik, "Cenozoic thermo-tectonic evolution of the Gangdese batholith constrained by low-temperature thermochronology," *Gondwana Research*, vol. 41, pp. 451–462, 2017.

[59] L. A. Kirstein, "Thermal evolution and exhumation of the Ladakh Batholith, northwest Himalaya, India," *Tectonophysics*, vol. 503, no. 3-4, pp. 222–233, 2011.

[60] P. van der Beek, J. van Melle, S. Guillot et al., "Eocene Tibetan plateau remnants preserved in the northwest Himalaya," *Nature Geoscience*, vol. 2, no. 5, pp. 364–368, 2009.

[61] P. J. Dézes, J. C. Vannay, A. Steck, F. Bussy, and M. Cosca, "Synorogenic extension: Quantitative constraints on the age and displacement of the Zanskar shear zone (northwest Himalaya)," *Geological Society of America Bulletin*, vol. 111, no. 3, pp. 364–374, 1999.

[62] K. E. Sundell, M. H. Taylor, R. H. Styron et al., "Evidence for constriction and Pliocene acceleration of east-west extension in the North Lunggar rift region of west central Tibet," *Tectonics*, vol. 32, no. 5, pp. 1454–1479, 2013.

[63] R. C. Thiede, J. R. N. Arrowsmith, B. Bookhagen, M. McWilliams, E. R. Sobel, and M. R. Strecker, "Dome formation and extension in the Tethyan Himalaya, Leo Pargil, northwest India," *Geological Society of America Bulletin*, vol. 118, no. 5-6, pp. 635–650, 2006.

[64] J.-C. Vannay, B. Grasemann, M. Rahn et al., "Miocene to Holocene exhumation of metamorphic crustal wedges in the NW Himalaya: evidence for tectonic extrusion coupled to fluvial erosion," *Tectonics*, vol. 23, no. 1, 2004.

[65] N. Buchs and J.-L. Epard, "Geology of the eastern part of the Tso Morari nappe, the Nidar Ophiolite and the surrounding tectonic units (NW Himalaya, India)," *Journal of Maps*, vol. 15, no. 2, pp. 38–48, 2019.

[66] R. Zhang, M. A. Murphy, T. J. Lapen, V. Sanchez, and M. Heizler, "Late Eocene crustal thickening followed by Early-Late Oligocene extension along the India-Asia suture zone: evidence for cyclicity in the Himalayan orogen," *Geosphere*, vol. 7, no. 5, pp. 1249–1268, 2011.

[67] A. Replumaz, A. M. Negredo, A. Villaseñor, and S. Guillot, "Indian continental subduction and slab break-off during Tertiary collision," *Terra Nova*, vol. 22, pp. 290–296, 2010.