GEOSCIENCES

Special Topic: The Tibetan Plateau

Paleoaltimetry reconstructions of the Tibetan Plateau: progress and contradictions

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ABSTRACT

Over the last two decades, many quantitative paleoaltimetry reconstructions of the Tibetan Plateau have been published, but they are still preliminary and controversial, although several approaches have been combined paleontology and geochemistry, including vertebrate, plant, and pollen fossils as well as oxygen, carbon, and hydrogen isotopes. The Tibetan Plateau is the youngest and highest plateau on Earth, and its paleoaltimetry reconstructions are crucial to interpret its geodynamic evolution and to understand the climatic changes in Asia. Uplift histories of the Tibetan Plateau based on different proxies differ considerably, and two viewpoints are pointedly opposing on the paleoaltimetry estimations of the Tibetan Plateau. One viewpoint is that the Tibetan Plateau did not strongly uplift to reach its modern elevation until the Late Miocene, but another one, mainly based on stable isotopes, argues that the Tibetan Plateau formed early during the Indo-Asian collision and reached its modern elevation in the Paleogene or by the Middle Miocene. With either a geochemical or paleontological approach, the present is used as the key to the past. However, there are great difficulties because modern processes of isotopic fractionation and species for creature distribution are not easily precisely determined. In addition, the climatic and environmental backgrounds of past geological times have massive differences from the present, and associated adjustments are influenced by many human factors. In the future work, the applications of multidisciplinary comprehensive methods and cross-checks of their results will be productive, and we look forward to achieving more reliable estimates for paleoelevations of the Tibetan Plateau.

Keywords: Tibetan Plateau, Cenozoic, paleoaltimetry, paleontology, stable isotope

INTRODUCTION

In the mid-19th Century, Scottish geologist, paleontologist, and botanist Hugh Falconer discussed the upheaval of the Tibetan Plateau. He reported some rhinocerotid fossils collected from the Zanda Basin north of the Niti Pass in Nagri, Tibet, China from the Tibetan merchants across the pass [1]. Modern rhinos still live in the Indian plains, so Falconer naturally considered that the rhinos from the Zanda Basin should also indicate low elevation as for living animals. The age of these rhinocerotid fossils meant that since several million years ago, the Himalayas have uplifted by more than 2000 m.

A century later, quantitative estimations for the Tibetan Plateau uplift were made again based on plant and mammalian fossils in the 1970s [2,3], but subsequent progress has been slow. For the recent decades only, the quantitative paleoaltimetry reconstructions of the Tibetan Plateau have developed gradually and become a research focus, since Garzione et al. [4] suggested the first empirical formula to estimate the paleoaltimetry of the Tibetan Plateau quantitatively using stable oxygen isotopes. At present, the quantitative paleoaltimetry study of the Tibetan Plateau is still preliminary, and several paleoaltimetry methods are used, including oxygen isotope, ∆47 paleotemperature, hydrogen isotope, paleontology, and paleoclimatology [5].

The Tibetan Plateau is the youngest and highest plateau on Earth, and its elevation reaches one-third of the height of the troposphere, with profound dynamic and thermal effects on atmospheric circulation and climate [6,7]. The uplift of the Tibetan Plateau was an important factor of global...
climate change during the late Cenozoic and strongly influenced the development of the Asian monsoon system [8,9]. Reconstructing the paleoelevation of the plateau can improve our understanding of the linkage between tectonics and long-term climate changes. However, there have been heated debates about the history and process of Tibetan Plateau uplift, especially over paleoelevations in different geological ages, because there are few direct indicators of paleotopography in the geological record [4,9,10–23].

Two viewpoints are pointedly opposed on the paleoaltimetry estimations of the Tibetan Plateau. One viewpoint is that the Tibetan Plateau did not strongly uplift to reach its modern elevation until the Late Miocene [2,21,24–32]. Another viewpoint, mainly based on stable isotopes, argues that the Tibetan Plateau formed at an early stage of the Indo-Asian collision and reached its modern elevation in the Paleogene or Middle Miocene [11,12,14–17,19,23,33,34].

Uplift histories of the Tibetan Plateau based on different proxies differ considerably. In the Gyirong Basin, for example, the Late Miocene paleoelevation based on stable carbon isotopes of tooth enamel of fossil herbivores would be about 2500 m [21] lower than that based on stable oxygen isotopes of lacustrine carbonates [17]. In the Namling Basin, the Middle Miocene paleoelevation based on the distribution of nearest living relatives (NLRs) of fossil plants and pollens is 1600 m [28] lower than that based on the climate leaf analysis multivariate program (CLAMP) [19], and 2200 m lower than that based on the stable oxygen isotope composition of pedogenic and early diagenetic carbonates [10]. In the Lunpola Basin, the Early Miocene paleoelevation based on the mammalian and pollen fossils [13,30] is over 1000 m lower than that based on stable oxygen isotopes of the paleosol carbonates and lacustrine limestones as well as the hydrogen isotope composition of n-alkanes from epicuticular plant waxes [16,34].

Whether geochemical methods or paleontological ones are employed, the present is used as the key to the past. However, there are great difficulties. For the isotopic fractionation and species distribution, modern processes are not easily determined precisely. In addition, the climatic and environmental backgrounds of the geological time have massive differences from the present, so associated adjustments have to be influenced by many human factors.

In order to summarize and review the recent advances of paleoaltimetry researches for the Tibetan Plateau, we discuss the main study areas (Fig. 1).
In each area, different results were usually obtained by different or the same methods and similar results were achieved by different methods. We hope that our comments are beneficial to the future work in the paleoaltimetry estimations of the Tibetan Plateau so that the uplift process can be reconstructed as accurately as possible.

**Himalayan Mountains**

**Middle Eocene: near sea level of the Himalayan Mountains**

A well-known fact is that the Himalayan Mountains began to uplift after the Indo-Asian collision. Although the knowledge of the initial timing between India and Asia collision was still controversial, most geological and geophysical evidences constrained it as 55 ± 10 Ma [35–40]. However, collision does not mean that the Himalayas had been uplifted. The timing of the uppermost marine successions disappearing was considered the true starting point for uplift of the Himalayan Mountains. Transition from marine to terrestrial successions at Zanskar area, India, western Himalaya syntaxis occurred at ~52 Ma, which implicates that this area began to uplift [41,42]. While the Himalaya foreland basin in Pakistan terminated its Balakot Group marine deposits at ~45 Ma, analogous to the Cuojiangding Group in the Zhongba area of Tibet. The continental sourced sandstones and conglomerates of the Jidula Formation firstly deposited in the Gangba and Dingri area after 50 Ma [43,44], representing the Himalayas began to uplift in this area. Summaries above different transition timing from marine to terrestrial deposits, most Himalayan area was close to or above sea-level land by 50 Ma.

**Middle-Late Miocene: near-present elevation of the Himalayan Mountains?**

In the Himalayan Range, a series of N-S trend grabens were developed during Middle-Late Miocene from west to east. The deposited sedimentary successions contain abundant carbonates and fossils, which can be used to estimate the paleoelevations of these basins by different paleoaltimetry methods.

**Zanda Basin**

The Zanda Basin is located in the drainage area of Langqên Zangbo (river, upper reaches of the Sutlej River) in Nagri, southwestern Tibet, China, with elevations of 3700—4500 m above sea level (a.s.l.). Tectonically, the Zanda Basin is located at the boundary between the Himalayan orogenic belt and the Lhasa terrace. The almost horizontal late Cenozoic strata of the basin have about 800 m maximum thickness [45].

Oxygen isotopic compositions of meteoric water ($\delta^{18}O_{MW}$) vary as a function of elevation, decreasing by global average values of about $-2.8 \%/\text{km}$ [46]. In the ideal case, the $\delta^{18}O$ value of surface water and minerals precipitated from that water reflect the average $\delta^{18}O_{MW}$ value of rainfall in the catchment, so oxygen isotopes can be used to reconstruct paleo-elevations [47,48].

Saylor et al. [18] examined the paleoelevation by analysis of well-preserved fossil mollusks and plant remains from the Late Miocene to the Early Pleistocene in the Zanda Basin. Based on $\delta^{18}O$, (carbonate) values from shell aragonite, they estimated that oxygen isotope ratios of Miocene–Pleistocene paleosurface water ($\delta^{18}O_{PSW}$) in this basin ranged from $-24.5$ to $-2.2\%$ (Vienna Standard Mean Ocean Water: VSMOW, the same below). The lowest of these calculated values are lower than $\delta^{18}O_{PSW}$ values from modern water in Zanda. The extremely low $\delta^{18}O_{PSW}$ values from fluvial mollusks and evaporatively elevated $\delta^{18}O_{PSW}$ values from lacustrine mollusks, show that the peaks surrounding the Zanda Basin were at elevations at least as high as, and possibly up to 1500 m higher than today. But a decrease in elevation since the Miocene is not specifically predicted by any existing mechanical models for the development of the Tibetan Plateau.

The Pliocene mammalian fauna of the Zanda Basin shows initiation of cold-adapted lineages that predate Ice Age megafauna, and the best examples include the earliest arctic fox (Vulpes quzhudingi; Fig. 2A), the Tibetan woolly rhino (Coelodonta thibetana; Fig. 2B), and the ancestral snow leopard (Panthera blytheae; Fig. 2C) [49–51]. Compared with their Late Pleistocene or living descendants, these cold-tolerant species became widespread during the subsequent Pleistocene Ice Age, and the Tibetan Plateau had thus become a cradle for Ice Age megafauna, which implied that the Zanda Basin reached its modern elevation with a mean annual temperature (MAT) close to 0°C [49,52–54].

Because both morphology and attachment impressions on fossilized bones can reflect muscular and ligamentous attachments, they can provide evidence for the type of locomotion that extinct animals used when they lived. A skeleton of the three-toed horse *Hipparion zandaense* discovered from the Zanda Basin provides an opportunity to reconstruct its locomotive function [55]. The anatomical features of *H. zandaense* indicate an ability to run fast and stand persistently, which is beneficial only on an
open habitat and reflects an adaptation for grassland instead of forest [56,57].

The Himalayas have appeared as a mountain range since the Miocene at least, with the appearance of vegetation vertical zones following thereafter so that the open landscape must be above the tree line [58]. In the Zanda area, the modern tree line is at an elevation of 3600 m between the closed forest and the open steppe [59]. When H. zandaense lived at 4.6 Ma, the mid-Pliocene global climate was about 2.5 °C warmer in temperature than today [60,61]. Assuming a temperature lapse rate of 0.65 °C/100 m applies to the past, the elevation of the tree line in the Zanda area was about 4000 m in that time. Because of the locality of the H. zandaense skeleton is situated at near 4000 m now, it demonstrates that the Zanda Basin had achieved an elevation comparable to its present-day elevation in the mid-Pliocene [55].

On the contrary, H. xizangense from the early Late Miocene (∼10 Ma [62]) the Bulong locality at a modern elevation of 4560 m in Biru County, Tibet has comparatively low cheek teeth, and its third metatarsals lack a facet for the fourth tarsal, both being the characters of woodland-forest horses. This is consistent with the ecosystem of the Bulong fauna and flora [63], indicating the elevation should be lower than 2500 m a.s.l. [55].

If the Zanda Basin was indeed 1500 m higher in the Mio-Pliocene than today [18], its MAT in the Mio-Pliocene would be about −10 °C (i.e. ∼10 °C lower than today) assuming no significant change in global temperature since the Mio-Pliocene and according to the present-day atmospheric lapse rate of 0.65 °C/100 m. That is, such low MAT seems unlikely given the abundance and diversity of fossil mammals, fish and shells found in the Late Miocene–Pliocene strata [52], and few modern types of vegetation can survive at an elevation above 5500 m a.s.l. [64].

In the late Cenozoic deposits of the Zanda Basin, fossil mammals are the most abundant in the middle section where most samples were collected [52]. Wang et al. [54] analyzed the carbon and oxygen isotopic compositions of structural carbonate in bioapatite of serial and bulk enamel samples from a diverse group of fossil and modern herbivores as well as the oxygen and hydrogen isotopic compositions of precipitation and stream waters from the basin. They estimated paleotemperatures using two methods: the carbonate clumped isotope thermometry [65] and a fossil-based oxygen isotope temperature proxy [7,8]. Their results supported the paleoelevation based on the paleoecology of the fossil mammals.

The δ13C values of enamel samples from modern wild Tibetan asses and domestic horses, cows, and goats in Zanda are −9.4 ± 1.8‰, which indicate a diet comprising predominantly C3 plants [9,66] and are consistent with the current dominance of

Figure 2. Fossils of the Pliocene cold-adapted mammals from the Zanda Basin in the Tibetan Plateau. (A) Vulpes qiuzhudingi, (A1) right hemimandible fragment in lingual view, (A2) right hemimandible fragment in buccal view, (A3) left hemimandible in lingual view, (A4) left hemimandible in buccal view [51]; (B) Coelodonta thibetana in lateral view, (B1) skull, (B2) mandible [49]; (C) skull of Panthera blytheae, (C1) dorsal view, (C2) lateral view, (C3) ventral view [50].
C3 vegetation in this area. The enamel δ13C values of the fossil horses, rhinos, deer, and bovids are $-9.6 \pm 0.8\%$, indicating that these ancient mammals, like modern herbivores in the area, also fed primarily on C3 vegetation and lived in an environment dominated by C3 plants. The lack of significant C4 plants in the basin suggests that the area had reached high elevations (>2500 m) [67] by at least the mid-Pliocene.

Paleotemperature estimations derived from a fossil bone-based oxygen isotope temperature proxy [7,8] as well as the carbonate clumped isotope thermometer [65] for the mid-Pliocene Zanda Basin are higher than the present-day MAT in the area. After accounting for the late Cenozoic global cooling, the paleotemperature estimation suggested that the paleoelevation of the Zanda Basin in the mid-Pliocene was similar to or slightly less than (~1000 m) its present-day elevation, which is consistent with the inference from the δ13C data [54]. On the other hand, Wang et al. [54] also indicated that this method has large uncertainties.

Clumped isotope thermometry uses measurements of the 13C–18O bond ordering in carbonates to constrain the temperature [T(Δ47)] and δ18O value of the water from which the carbonate originated. These data can be used to infer paleoelevation by exploiting the systematic decrease of surface temperature and the δ18Ow with elevation, provided samples record original depositional conditions and appropriate context exists for interpreting T(Δ47) and δ18O values [65,68,69]. Huntington et al. [70] revisited with carbonate clumped isotope thermometry the conventional stable isotope studies of samples recording primary δ18Ow from the Zanda Basin. Their conclusion is that the Late Miocene–Pliocene Zanda Basin was colder and therefore higher than the present.

Gyirong Basin

The Gyirong Basin is a late Cenozoic intermontane basin in the central Himalayas with an area of about 200 km². This basin is bounded by an active fault on the east and lies at elevations of 4100—4500 m a.s.l. It was suggested that the Gyirong Basin was originally an east–west trending graben formed by the north–south extension related to the south Tibetan detachment system and was later deformed by the north–south trending normal fault to form the present north trending valley [71]. A thick sequence of lacustrine and fluvial sediments began to accumulate in the basin at 7.2 Ma [72].

In the expedition to the Tibetan Plateau organized by the Chinese Academy of Sciences during the 1970s, a Late Miocene Hipparion fauna was found in the Gyirong Basin [73]. The fossil locality at Woma, Gyirong has a modern elevation of 4384 m, and the ecological features of this fauna indicate a mix of forest and grassland mammals. The Gyirong Hipparion fauna is quite different from the Siwalik Hipparion fauna in South Asia, which implies that the Himalayas had become an obvious barrier for faunal migrations since the Late Miocene [73,74].

The paleoelevation of the Gyirong Basin was quantitatively determined to be 5850 $+1410 \text{ m/}−730 \text{ m}$ a.s.l. by Rowley et al. [17] based on the paleo δ18O measured from the most negative lacustrine carbonate δ18Oc value of $−21.5\%$ reported by Wang et al. [75]. The underlying assumptions of this paleoelevation estimation include the following. (i) The carbonate was formed in the Late Miocene at 2°C, same as today’s MAT. (ii) The most negative δ18Oc value records the δ18Ow. (iii) The paleo δ18Oc in the Late Miocene at Bakiya Khola, Nepal, which was considered as a low-elevation reference site near the moisture source, was the same as the δ18Ow of soil water calculated from the δ18Oc values of paleosol carbonates and the assumed paleotemperature. And (iv) the variation in δ18O in water elevation is predictable using a one-dimensional Rayleigh-type fractionation model.

On the other hand, this estimation clashes with the presence of significant C4 biomass in the area as indicated by the carbon isotope data from fossil teeth and paleosols [4,21], also inconsistent with clay mineralogy and pollen and mammalian fossil evidence [20,73,76,77].

Although precipitation is in general more depleted in 18O at high elevations than at low elevations, precipitation in the Himalayan-Tibetan region displays large δ18Ow variations on various spatial and temporal scales, which cannot be explained by elevation alone [9]. An altimetry estimation of an area during a geologic period using an empirical δ18Ow versus elevation relationship or a Rayleigh-type fractionation model would be reliable only if the climate conditions of the study area during the geologic time of interest were the same as today. This assumption is clearly invalid for the Gyirong area where various lines of evidence indicate a warm and humid climate in the Late Miocene [20,21,73,76], very different from today’s cold and dry environment.

The δ13C value of tooth enamel of the modern horse, yak, and goat in the Gyirong Basin is between −9.0 and $−14.2\%$, with a mean of $−12.2 \pm 1.5\%$, indicating a pure C3 diet, which is identical with the dominant modern C3 vegetation [78]. On the other hand, the δ13C value of tooth enamel of the Late Miocene Hipparion fossils at 7 Ma in the Gyirong Basin is between $−2.4$ and $−8.0\%$, with a mean of
−6.0 ± 1.1‰, indicating their mix diet of C₃ and C₄ in which C₄ plants accounted for 30~70%; the enamel δ¹³C value of the rhinocerotid *Chilotherium* fossils is −7.9 ± 0.1‰, also indicating a C₃/C₄ mix diet in which C₄ plants account for 30%. This result shows that the Late Miocene ecological environment was a sparse woodland, consistent with the evidence based on spore and pollen analysis [21].

C₄ plants have an advantage over C₃ plants in conditions of higher temperature, better light, and plentiful moisture. The distribution of modern C₄ plants is controlled by temperature, seasonal precipitation and elevation, widespread in tropical and subtropical areas with low elevation under 2500 m a.s.l., but scarce and even absent in areas with circumpolar latitude or high elevation above 3000 m and areas characterized by winter precipitation. The stable carbon isotopes demonstrate that C₄ plants were present in the Gyirong Basin and dominant in the ecosystem during the Late Miocene, indicating climate with higher temperature and environment with lower elevation in this area. The mean δ¹⁸O value of tooth enamel from fossil horses in the Gyirong Basin (−17.0 ± 1.5‰) is significantly higher than that of modern horses (−19.1 ± 0.6‰) in this basin, which appears to be consistent with the interpretation of a lower elevation in the Late Miocene from the stable carbon isotopes.

In temperate environments today, C₄ grasses are abundant in warm months below 1500 m elevation, but they either disappear completely or account for negligible amounts of the total biomass at elevations above 2500 m [79]. Even in the tropics, grasslands above 3000 m elevation are dominated by C₃ grasses with no or insignificant amounts of C₄ grasses [80,81]. Assuming that the atmospheric pCO₂ level in the Late Miocene was about the same as it is today [82] and that the average temperature lapse rate of 0.65°C/100 m elevation applies to the past, 6°C warmer temperatures in the Late Miocene [83] could extend the upper range of C₄-dominated grasslands to 2423 m at mid-latitudes and to 2923 m in the tropics. If the temperature was 9°C warmer (extreme scenario) in the Late Miocene [83], the upper range of C₄-dominated grasslands could extend to 2885 m at mid-latitudes and to 3385 m in the tropics. As a result, the carbon isotopes indicate that the elevation of the Gyirong Basin must have been lower than ~2900–3400 m, most likely lower than ~2400–2900 m [21].

Wang et al. [22] presented high-resolution oxygen and carbon isotope data at 0.4–0.5 m sampling interval in the Gyirong Basin. They applied a novel approach [8] that combines in vivo oxygen isotope signatures of fossil enamel with diagnostic oxygen isotope compositions of fossil bones. Studies have shown that the enamel δ¹⁸O values of obligate drinkers are strongly correlated with the δ¹⁸O_mw [84,85]. Using the enamel–water δ¹⁸O relationship of Kohn and Cerling [84], the estimated paleo δ¹⁸O_mw values in the Late Miocene are −19.6 ± 1.5‰.

The average δ¹⁸O value of lacustrine carbonates at 6.9–7.1 Ma is −15.2 ± 2.0‰ in Gyirong. Using the paleotemperature estimated above from fossils, they calculated the average δ¹⁸O_mw value of lake water in equilibrium with these lacustrine carbonates to be −14 ± 2‰. Paleotemperatures are 21 ± 6°C at ~7 Ma, which is ~19 ± 6°C higher than the present-day MAT in the same area. If the present-day atmospheric lapse rate of 0.65°C/100 m is used, the estimated temperature change since the latest Miocene would correspond to an elevation change of 2154 ± 923 m, implying that the paleolake basin was about 2230 ± 923 m a.s.l. at ~7 Ma. This temperature-based paleoaltimetry estimation is consistent with the carbon isotope data [21], indicating that the paleo-Woma lake basin in the Gyirong area and possibly a significant portion of the central Himalaya were at a much lower elevation than today.

**Mount Qomolangma area**

In the absence of fossil surface deposits such as palaeosols, volcanic ashes, or lake sediments, Gebelin et al. [86] conducted stable isotope paleo-elevation based on the hydrogen isotope ratios (δD) of hydrous minerals that were deformed in the South Tibetan detachment (STD) shear zone during the late Early Miocene. These minerals exchanged isotopically at high temperature with meteoric water (δDw = −156 ± 5‰) that originated as high-elevation precipitation and infiltrated the crustal hydrologic system at the time of detachment activity. When compared to age-equivalent near sea-level foreland oxygen isotope (δ¹⁸O) paleo records (δ¹⁸Ow = −5.8 ± 1.0‰) [87,88], the difference in δ¹⁸O_w is consistent with mean elevations of ≥5000 m for the Mount Qomolangma area. Mean elevations similar to modern suggest that an early Himalayan rain shadow may have influenced the late Early Miocene climatic and rainfall history to the north of the Himalayan chain.

They collected oriented samples from the STD in the underlying mylonitic footwall in the Rongbuk Valley, north of Mount Qomolangma. In this area with present day minimum and maximum elevations of 4553 and 5982 m, the STD consists of the upper (brittle) Qomolangma detachment and the lower (ductile) Lhotse detachment shear zone that merge northward and together separate upper plate, non-metamorphosed Ordovician limestone...
from high-grade metamorphic rocks and syntectonic leucogranite below. Based on radiometric dating of mylonitic and undeformed leucogranite, the maximum age of ductile shearing is \( \sim 17 \text{ Ma} \), while brittle faulting on the Qomolangma detachment was likely younger than \( 16 \text{ Ma} \) [89]; in the following, Gebelin et al. [86] refer to the timing of shearing on this part of the STD as late Early Miocene.

The relative difference in \( \delta^{18} \text{O} \) between late Early Miocene meteoric water in the low-elevation Siwalik foreland basin (\( \delta^{18} \text{O}_w = -5.8 \pm 1.0 \%) \) based on \( \delta^{18} \text{O} \) of pedogenic carbonate and MATs of [90]) and the high-elevation precipitation that infiltrated the STD (\( \delta^{18} \text{O}_w = -20 \pm 1 \% \) to \(-21 \pm 1 \% \) based on \( \delta D \) of biotite and hornblende, respectively) yields \( \Delta \delta^{18} \text{O}_w = 14.2 \pm 1.0 \% \) (calcite–biotite) and 15.0 \( \pm 1.0 \% \) (calcite–hornblende). Because stable isotopes in precipitation systematically scale with elevation (\( \sim 2.8 \%/\text{km} \) in \( \delta^{18} \text{O} \) or \( \sim 22 \%/\text{km} \) in \( \delta D \) [15,46]), this Miocene \( \Delta \delta^{18} \text{O}_w \) between Rongbuk Valley (STD) and the Siwalik basins is consistent with an elevation difference of \( \sim 5100 \pm 400 \text{ m} \) and \( \sim 5400 \pm 350 \text{ m} \), respectively, for the two mineral complexes. This would put the Mount Qomolangma area at \( \sim 6000 \text{ m} \).

**DIFFERENT UPLIFT HISTORIES OF THE LHASA TERRANE**

Until \( \sim 110 \text{ Ma} \), the widely distributed marine Taka and Langshan formations in the north of Lhasa terrane imply that most of the area was still below sea level. The Gangdese magmatic arc may have been no more than 200 m in elevation, displaying a normal continental crust [91]. Hereafter, abundant island arc volcanic rocks permeated the entire Lhasa terrane during 110–80 Ma, which may be the results of the New Tethyan lithosphere sustained subduction beneath the Lhasa terrane. This tectonic dynamics would cause to rapidly shorten and thickened the Lhasa lithosphere and also contributed rapid surface uplift during Late Cretaceous. However, when present heights of the Gangdese and northern Lhasa terrane were obtained still needs estimation by quantitative paleoelevations.

From the paleoelevations of the basins from the Lhasa terrane, a sketch of the different uplift histories of the northern and southern Lhasa terrane becomes gradually clear (Fig. 3).

**Andean-type Gangdese Mountains before Indo-Asian collision**

As one the most prominent tectonic unit of the south margin of the Eurasian plate, Gangdese Mountain Range elongate more than 4000 km from the Ladakh area to the west to Sang plateau to the east and are mainly composed of the Late Triassic to Eocene magmatic rocks, striking with link to the northward subduction of the New Tethyan lithosphere and collision between India and Eurasia.

Although the crustal deformation, surface denudation and sedimentary sources reveal the Gangdese arc underwent significant uplift, they still cannot depict whether a high-elevation Andean-type mountain range stood at the south margin of the Lhasa terrane. The Linzhou Basin, Namling Basin, and Kailas Conglomerate provide windows to probe this elevation history (Fig. 3).

**Linzhou Basin**

The Linzhou Basin is located in the Gangdese arc at the southern margin of the Lhasa terrane. In this basin, the Linzizong Group was deposited between 69 and 43 Ma [92,93] and unconformably overlies Cretaceous strata. The total thickness of the Linzizong Group in the Linzhou Basin is \( \sim 3500 \text{ m} \) with volcanic-bearing strata and constitutes three units from lower to upper: the Dianzhong, Nianbo, and Pana formations [92].

Ding et al. [14] presented oxygen and carbon isotopic compositions from well-dated Paleogene pelosols, lacustrine calcareous carbonates, and marls from the Nianbo (60–54 Ma) and upper Pana formations (51–48 Ma). The sediments of the Nianbo Formation, \( > 180 \text{ m} \) thick, were deposited in alluvial fans, braided rivers, fan deltas, in nearshore to offshore lacustrine settings, whereas those of the upper Pana Formation \( > 100 \text{ m} \) thick are comprised predominantly of proximal alluvial fan and braided river deposits.

Correlations between their lithofacies and stable isotopic compositions suggest that the Linzhou Basin was principally a hydrologically open environment. It should be stressed that the \( \delta^{18} \text{O}_w \) and \( \delta^{13} \text{C} \) values from the Nianbo and upper Pana formations have not yet been reset using late-stage diagenesis based on petrographic examination, fossil ostracode oxygen isotopes, and stratigraphic tectonic deformation.

The paleoelevations were reconstructed using corrected, most negative \( \delta^{18} \text{O}_{\text{pow}} \) values. Based on the reconstructed paleotemperatures of the global sea surface [94] and the Namling Basin [19], they conservatively use the average paleotemperatures of 18.5 \( \pm 8 \text{ C} \) and 13.8 \( \pm 8 \text{ C} \) to reconstruct the \( \delta^{18} \text{O}_{\text{pow}} \) values when the Nianbo and upper Pana formations deposited, respectively. They assumed that the most negative paleosol \( \delta^{13} \text{O} \) value of \(-19.2 \%) \) in the Nianbo Formation reflects the surface...
was also corrected up to $-400$ m using $3.4 \pm 0.5$ value of

values can be used to constrain the oxygen isotope

tions gives an MAT of $8.1 \pm 2.3^\circ C$ with a CMMT of $-4.19 \pm 3.7^\circ C$. The modest CMMTs suggest that
either data set is appropriate for analyzing the Namling

Figures 3. Comparison of paleoaltimetry histories between Gangdese, southern Lhasa terrane and northern Lhasa terrane during the Conozoic. The Gangdese arc at southern Lhasa terrane and Qiangtang terrane has kept the elevation above 4500 m since India–Asia collision [10,14,19,23,102], while the northern Lhasa basins underwent rapid erosion and uplift from a low-elevation peneplain [131,132] and finally obtained present elevation by Late Eocene–Early Oligocene [12,16]. However, data of the animal and pollen fossils (Triangles) indicate the Lunpola Basin was only 3000 m in elevation during Early Miocene [13,30]. The error bars represent the time and elevation uncertainties.

water surrounding the paleobasin catchments. The $\delta^{18}O_{psw}$ was also corrected up to $-16.9^\circ /oo$ between 60 and 54 Ma and $-15.0^\circ /oo$ between 51 and 48 Ma, respectively.

In conclusion, they implied that the Linzhou area had attained an elevation of $4500 \pm 400$ m using the thermodynamic relationship [17,95] during the period of the Indo-Asian collision, i.e. it achieved a near-present elevation, and may have formed an Andean-type mountain range stretching across the Gangdese arc before that collision (Fig. 4).

Namling Basin

The Namling Basin in southern Tibet comprises Neogene lava flows and shales interbedded with tuffaceous material unconformably overlying an Upper Cretaceous/Eocene volcanic basement. The western side of the basin has a series of steeply dipping, ash-rich, lacustrine sediments, within which two fossil leaf sites have been sampled [96]. The lower site lies at 4300 m and the upper site at 4600 m beneath capping lava flows at $>4800$ m [19]. The flora is dominated by Quercus sect. Heterobalanus [96], with an age of about 15 Ma based on $^{40}\text{Ar}^{39}\text{Ar}$ dating of the volcanicogen sediment lying immediately below the fossiliferous bed. Taxa of the floral assemblages showed no sign of biodegradation, which suggests that the overall species composition is not biased because of differential species decay.

Spicer et al. [19] exploit the empirical relationship between leaf physiology and properties of the atmosphere related to elevation by using the CLAMP [97,98]. Their specimens were categorized into 35 morphotypes using leaf architecture and venation characteristics and were subjected to a CLAMP analysis. CLAMP values for the Namling fossils yield an MAT of $6.8 \pm 3.4^\circ C$ and a cold month mean temperature (CMMT) of $-6.2 \pm 5^\circ C$ using a data set containing the so-called ‘alpine nest’ samples (Physg3ar). An alternative data set (Physg3br) that lacks samples from cold environments gives an MAT of $8.1 \pm 2.3^\circ C$ with a CMMT of $-4.19 \pm 3.7^\circ C$. The modest CMMTs suggest that either data set is appropriate for analyzing the Namling fossils.

Spicer et al. [19] considered the variation of lapse rates in mountainous terrain [99], and they used moist static energy (MSE) indirectly. To determine elevation, the method requires an estimate of MSE at a known elevation, and they had to calculate the Miocene MSE using the Hadley Centre numerical climate model. They also admitted that MSE is a relatively robust model variable, despite many uncertainties. As a result, they reconstructed the elevation of the Namling Basin to be $4689 \pm 895$ m or $4638 \pm 847$ m at 15 Ma, which is comparable to the present-day elevation of 4600 m. Therefore they concluded that the elevation of the southern Tibetan Plateau probably has remained unchanged for the past 15 Ma.

Currie et al. [10] used the stable isotope composition of pedogenic and early diagenetic carbonates from the Namling Basin to make model estimations of the paleoelevation of the Middle Miocene. Their samples were collected from the argillic and calcic horizons in the paleosols, which are interpreted to have formed as a result of translocation of authigenic clays and leaching of calcium carbonate from upper soil horizons by meteoric water during pedogenesis. Pedogenic calcium carbonate nodules have average $\delta^{18}O$ values of $-19.6^\circ/oo$, whereas nodular lacustrine dolomites range in composition from $-7.6$ to $-5.5^\circ/oo$. The most negative of the carbonate isotope values can be used to constrain the oxygen isotope composition of paleoprecipitation. Model results indicate that the Namling Basin achieved elevations of $\sim5200 + 1370/ -605$ m by at least 15 Ma, which is close to the floral-based estimate of [19].

By using the fractionation factors for calcite-water and dolomite-water determined by Friedman and O’Neil [100] and the paleo MAT of $6.8^\circ C \pm 3.4^\circ C$ determined by the Spicer et al. [19] leaf-margin analysis of the Namling flora, Currie et al. [10] estimated the paleo $\delta^{18}O_{nw}$. The pedogenic calcite nodules described here precipitated from
meteoric water with a $\delta^{18}$O$_{mw}$ value of $-21.7 \pm 1.7\%$o. Similarly, the most negative of the nodular dolomites precipitated from groundwater that had a $\delta^{18}$O$_w$ value of $-15.6 \pm 3.6\%$o.

For the low-elevation reference, Currie et al. [10] used $\delta^{18}$O$_c$ values of pedogenic carbonates from 17–14 Ma Siwalik Group sediments that yield mean values of $-9.5 \pm 1.3\%$o [101]. Assuming a MAT of 25°C, they estimated a low-elevation $\delta^{18}$O$_{mw}$ value of $-7.2\%$o for Middle Miocene paleoprecipitation. The effect of normalizing their estimation of $\delta^{18}$O$_{mw}$ with the Middle Miocene Siwalik value relative to the $-6\%$o value of today is to reduce predicted elevations. Given these values, their best estimate
of Middle Miocene $\Delta(\delta^{18}O_{\text{sw}})$ derived from pedogenic calcite nodules from the Himalayan foreland and the Namling Basin is $-14.5 \pm 2.1\%$. The uncertainty reflects combined contributions of the low-elevation $\delta^{18}O_{\text{sw}}$ value ($\pm 1.3\%$) and the temperature at which carbonate and water were in equilibrium during carbonate precipitation ($\pm 1.7\%$).

Contrary to the conclusion of Spicer et al. [19] and Currie et al. [10], Zhou et al. [28] suggested that the lower and upper sites are so different that they cannot be considered the same fossil assemblage. The lower site has a deciduous broad-leaved forest dominated by *Populus* and *Betula*, and its paleoelevation was 2500—3000 m, associated with the coexistence of NLRs. The upper site dominant genera *Queues*, *Rhododendron*, and *Thermopsis* coexist today at 2800—3000 m.

Zhou et al. [28] indicated that Spicer et al.’s conclusion is questionable in three areas as follows. (i) Elevation was estimated using CLAMP, which provides a subjective determination of the same leaf architecture by different users, and using inaccurate descriptions of leaf architecture. (ii) The sea level enthalpy in MSE was uncertain because it was taken from a Miocene climate model. The data set Physg3ar for CLAMP was established mainly based on climate parameters and the vegetation of North America [97,102], which differs from Asia in both the types of species and vegetation. And (iii) Spicer et al. [19] omitted comparisons with NLRs, and this is inappropriate because, except for *Thermopsis* and *Rhododendron*, non-NLR Namling fossil genera live today at Namling’s present elevation of 4800 m.

**Kailas Basin**

The Kailash Basin developed during Late Oligocene–Early Miocene with a >2.5 km thick Kailas Conglomerate [103]. Provenance and paleocurrent data indicate that the bulk of the Kailas Conglomerate was derived from the northerly Gangdese magmatic arc. The upper Kailas Conglomerate contains nodular red beds. These paleosol carbonates yielded $\delta^{18}O$ values ranging mostly between $-15$ and $-18\%$, which are lower $\sim 2-3\%$ than the most negative modern soil carbonate values of $-13$ to $-14.5\%$ in the Kailas and Nima areas of central Tibet. If the modest isotopic effects of $\sim 2-3\%$ of warmer, and largely ice-free condition in the Early Miocene were corrected to the paleosol carbonate values, they are comparable to the modern soil values. Thus, the oxygen isotope data suggest that the paleoelevation of the Kailas Formation during deposition of the red bed member was essentially the same as its modern elevation [103].

**High or low elevation of the northern Lhasa terrane?**

**Nima Basin**

The Nima Basin lies near the geographic center of the Tibetan Plateau, within a $\sim 50$ km radius of the county town of Nima, $\sim 450$ km northwest of Lhasa. The area is in the north-central part of the Lhasa terrane and in the southern part of the Bangong suture zone which separates the Qiangtang and Lhasa terranes. Over a distance of $\sim 150$ km east–west along the suture zone lie widespread outcrops of nonmarine strata of Middle Cretaceous to middle Cenozoic age [104,105].

DeCelles et al. [12] used stable isotope analyses of modern and accurately dated ancient paleosol carbonate in the Nima Basin to point to an arid climate and high paleoelevation (4500–5000 m, comparable to today’s setting) by 26 Ma. Oxygen isotope values of ancient soil carbonate are both very negative and indistinguishable, after modest corrections for changes in global climate, from the lowest (least evaporated) oxygen isotope values of modern soil carbonates in the area. Substantial enrichments in $O^{18}$ in paleolacustrine carbonates, as well as high carbon isotope values from paleosol carbonates, indicate considerable lake evaporation and low soil respiration rates, respectively, and both are consistent with the present arid climate of the Nima area. Oxygen isotope ratios in unaltered aquatic fossils also support the conclusion that the Nima Basin has been at high elevation since the Late Oligocene (Fig. 3). However, their conclusion draws from an admittedly limited data set.

On the other hand, $T(\Delta47)$ values of the marl samples analyzed by DeCelles et al. [12] from the southern part of the Nima Basin exceed Earth-surface temperatures, indicating that the samples have been diagenetically altered. Maximum burial temperatures were not high enough to cause solid-state C–O bond bond reordering. Instead, the elevated $T(\Delta47)$ and water $\delta^{18}O$ values are consistent with recrystallization of the samples in a rock-buffered system [70].

Wang and Wu [106] described a new Oligocene-aged genus and species of cyprinid fish, *Tchunglinius tchangii*, from the Nima Basin. The fish-bearing strata belong to the lower part of the middle Cenozoic non-marine deposits in the south Nima area, with ages of 23.5–26.0 Ma, Late Oligocene [12,105]. The new genus is assigned to the subfamily Cyprininae, and *T. tchangii* is closely related to recent small-bodied South Asian cyprinines such as *Puntius* in small body size. Therefore, the fossil reflects the Paleogene tropical–subtropical lowland fish fauna present there before the uplift of the Tibetan Plateau.
Tchunglinius tchangii has a smaller and shorter body with fewer vertebrae than other Tibetan fishes. Cyprinids living in cold water have more vertebrae than their congeners living in warm water [107]. Focusing on the recent fish fauna of the Tibetan Plateau and its surrounding area, cyprinid species living at high elevation have more vertebrae than those at low elevation. For example, Puntius, Systemus and Cosmochilus living in South Asia tropic–subtropic lowlands have 26–34 vertebrae, Acrossocheilus, Neolissochilus, Onychostoma, and Percocypris living in moderate elevations have 38–48 vertebrae, and Schizothorax, Ptychobarbus, Gymnodiptychus, Gymnocypris, Schizopygopsis, Choanchia, and Platypo- 
don living at high elevation and endemic to the Tibetan Plateau have 44–53 vertebrae [108, 109]. As to the fossil, T. tchangii with 33 vertebrae represents a Paleogene fish in tropic–subtropic lowland warm water, similar to recent South Asia before Tibetan Plateau uplift, while Plesiocichilus macrocephalus from the Lunpola Basin with 46–48 vertebrae below represents early Neogene fish in at least moderate elevation cool water after the Tibetan Plateau began to uplift [106].

Lunpola Basin

The Lunpola Basin is located on the southern and northern sides of the boundary between Baingoin and Shuanghu counties in northern Tibet. The basin has well-developed Cenozoic strata and an average elevation of about 4700 m a.s.l. The total thickness of Cenozoic deposits in the Lunpola Basin is over 4000 m, and consists of the lower Niubao Formation and the upper Dingqing Formation. Estimates of the paleoelevation of the Lunpola Basin have been very different. During the deposition of the Dingqing Formation, the lowest estimate has been reported to be about 1000 m [110], but the highest published estimate is 4900 m [34] (Fig. 3).

Rowley and Currie [16] used oxygen isotope paleoelevation of the paleosol carbonates and lacustrine limestones of the Lunpola Basin to reconstruct the elevation history of this area. They established a model of the relationship between elevation and $\Delta(\delta^{18}O_{mw})$ between sea level and elevation [17]. Their data from the Lunpola Basin indicate that the surface of the Tibetan Plateau has been at an elevation of more than 4000 m for at least 35 Ma.

Their paleoaltimetry estimation requires the identification of low-elevation reference sites. Ideally, these would be coeval, unequivocally low-elevation sites along the mean storm trajectories with which to compare data from the Lunpola Basin. Unfortunately, there are no data from the region to the south of the appropriate ages for the sections described above. Instead Rowley and Currie [16] used modern observations to place bounds on likely compositions of low latitude and low-elevation precipitation, and compare these with general circulation model-based estimation incorporating isotopes.

Oxygen isotopic compositions of the lacustrine carbonate ($\delta^{18}O_c$) of the Dingqing Formation range from $-1.3$ to $-14.6\%$ (PDB). This reported uncertainty reflects most importantly uncertainty in the temperature at which carbonate and water equilibrated, and the corresponding uncertainty in the estimation of the mean elevation; it also includes model-related uncertainties reflecting variations in temperature and relative humidity of low-latitude and low-elevation air masses [17].

In modern plants, the $\delta^D$ of leafwaxes reflects the $\delta^D$ of their growth water [111, 112]; therefore, $\delta^D$ measurements on ancient molecules are a potential means to estimate past water $\delta^D$ values and, ultimately, paleoelevation. Plant-wax $\delta^D$ values provide estimation for precipitation $\delta^D$ that is less changed by evaporation than lake calcite $\delta^{18}O$. Polissar et al. [34] measured the hydrogen isotope composition of n-alkanes from epicuticular plant waxes preserved in lacustrine deposits to reconstruct the $\delta^D$ of precipitation in Cenozoic basins that have been elevated as part of the Tibetan Plateau. n-Alkanes $\delta^D$- and carbonate $\delta^{18}O$-inferred water compositions from the Eocene–Miocene Lunpola Basin and Miocene Hoh-Xil Basin plot near enriched values relative to the global meteoric water line, as expected for evaporative lakewater and leafwater systems that have the same precipitation source. n-Alkanes $\delta^D$-based water compositions are nearly identical to the minimum carbonate $\delta^{18}O$-based values, demonstrating that plant-wax $\delta^D$ is minimally affected by evaporation compared to lacustrine calcite $\delta^{18}O$. This agreement strongly supports the presence of similar precipitation isotopic compositions in both archives despite different isotope systems, source water reservoirs, archive materials, modes of incorporation, and diageneric processes.

In the Lunpola Basin, the paired carbonate $\delta^{18}O$- and lipid deuterium-based estimates of paleoelevations from the Eocene Niubao Formation range from 3600 to 4100 m, and Miocene Dingqing Formation elevations range from 4500 to 4900 m. This result is close to the carbonate $\delta^{18}O$-based paleoelevations [16]. If atmospheric circulation was similar to the modern, however, paleoelevation values are 4300–5400 m and 7700–8000 m for the Niubao and Dingqing formations, respectively.

The zoogeography of the living schizothoracin fish has been studied and the evolution of the group deduced [113, 114]. They were considered to have
coevolved in correlation with the uplift of the Tibetan Plateau and were divided into three grades: primitive, specialized, and highly specialized. The division of the three grades was based on the extent of modifications of their scales, barbells, and rows of pharyngeal teeth, and their distribution at three successive elevations for which water temperature and precipitation decrease, and solar radiation and evaporation increase. The primitive grade inhabits water systems at an elevation from 1250 to 2500 m a.s.l., the specialized grade from 2750 to 3750 m, and the highly specialized grade from 3750 to 4750 m.

The fossil schizothoracin *Plesiachizothorax macrocephalus* from the Dingqing Formation falls in the distribution area of the modern highly specialized grade, but it has three rows of pharyngeal teeth and fewer vertebrae, and should be referred to the primitive or middle grade. All the fossil localities yielding primitive grade schizothoracins are situated well above the elevations tolerated by the living schizothoracins of the primitive grade. Thus, Wang and Chang [115] suggested that the elevations of the Lunpola Basin and other parts of the Tibetan Plateau would have been considerably lower during the period from the Oligocene to Pliocene than it is today.

A distal extremity of rhinocerotid humerus, found in the Dingqing Formation at a modern elevation of 4624 m, is almost the same in morphology as that of *Plesiaceratherium gracile* of the late Early Miocene Shanwang fauna from Shandong Province. The Shanwang fauna was dominated by mammals living in a forest edge and swampy area, indicating that the ecosystem was a subtropical or warm temperate forest [13]. The melanic deposits are well developed in the Dingqing Formation, which indicates a humid climate or environment as well [116,117].

The distribution of animals and plants on the southern slope of the Himalayas has distinct vertical zones, and the upper limit of the evergreen broad-leaved forest is 2500 m a.s.l. with a warm and humid climate, an annual rainfall of about 2000 mm [118], which is similar to the *Plesiaceratherium* habitat in Shanwang and Lunpola. The modern temperature is 4°C lower than that hypothesized for *Plesiaceratherium* in the late Early Miocene [119] when the evergreen broad-leaved forest would have been at a maximum elevation of 3115 m according to a temperature lapse rate of 0.65°C/100 m. The modern rhinos still live in the south side of the Tibetan Plateau where the highest distribution of the Javan rhino (*Rhinoceros sondaicus*) was recorded at 2000 m a.s.l. [120]. Based on the temperature increase of 4°C, the maximum paleoelevation of *Plesiaceratherium*’s habitat in the Lunpola Basin should be close to 3000 m [13].

Sun *et al.* [30] used a high resolution palynological record of the Dingqing Formation to indicate that the vegetation types during the latest Oligocene—earliest Miocene were dominated by mixed coniferous—broad-leaved forests being different from the modern steppe vegetation in the Lunpola Basin. The coexistence approach (CoA) was used in the fossil pollen records to reconstruct ranges, which are tolerated by all of the NLRs of these fossil plant taxa. The maximum overlap of the environmental tolerances of all the NLRs is then regarded as being indicative of the most likely paleoenvironment [122]. In Sun *et al.* [30], the temperature difference between the geological past and the present, used for correcting paleoelevation, is not deduced by the CoA method but by general circulation modeling results with considerations of the past boundary conditions (e.g., paleogeographic position, CO₂ concentrations, and CH₄ values).

In using the CoA, Sun *et al.* [30] compiled the elevation ranges of modern Tibetan vegetation equivalents of fossil pollen assemblages in the latest Oligocene—earliest Miocene. The maximum overlap of all the NLRs provides a coexistence elevation of 2400–2600 m a.s.l., averaged at 2500 m. According to a temperature curve from the Mg/Ca ratio of benthic foraminiferal calcite [123,124], they used the averaged lapse rate (0.579°C/100 m) and the temperature difference (4°C) during Eocene and Middle Miocene estimated by Song *et al.* [122] to yield a corrected elevation factor of 690 m, allowing the raw CoA elevation of 2400–2600 m to be increased by 690 m, so a maximum paleoelevation is 3190 ± 100 m, which is 1500–2000 m lower than the previous oxygen isotope paleoelevation in the Lunpola Basin [16,34].

For oxygen isotope paleoelevations, Sun *et al.* [30] considered that such studies have great uncertainties due to: (i) the commonly used authigenic mineral is carbonate, which is an unstable mineral under surface geochemical environment, so both the diagenesis and the climate-induced chemical weathering can alter the isotopic compositions of carbonate; and (ii) the Tertiary hydrological conditions, temperature and water vapor sources must be different compared from the present scenario; therefore, the present stable isotopic altimeter cannot be directly used for the geological past.

Jia *et al.* [125] also used the two proxies of carbon and hydrogen isotopic compositions of leaf wax n-alkanes from the Dingqing Formation to reconstruct paleoelevation of the Lunpola Basin. According to their analysis, values of δ¹³C and δD of C29 n-alkane were −29.8 ± 0.7‰ (n = 38) and −188 ± 10‰ (n = 22), respectively. Their measurements of δD of the isoprenoid compound phytane indicated that...
the original n-alkane δD signal was preserved during sedimentation. The time-series records exhibit clearly opposite trends from 25.5 to 21.6 Ma, i.e. a decreasing trend in δ13C29 but an increasing trend in δD29. After that until 20.4 Ma (below the top two samples), δ13C29 and δD29 turned to increase and decrease, respectively.

The δ13Ccom (the δ13C of the soil organic matter) of average C1 species-level response to elevation is $1.2 \pm 0.9 \%$/km with ranges from 1.1 to 1.3 \%$/km from numerous mountain ranges of the globe (e.g. [126–129]). By applying this elevation δ13C gradient to the observed δ13C differences between the Lunpola Basin and the Siwalik area [130], a mean paleoelevation of $\sim 3040 \pm 560$ m for the Lunpola Basin can be inferred. The relative difference in δ18Ow between the areas yielded a mean Δδ18Ow value of $-5.1 \%$ ranging from $-7.8$ to $-3.2 \%$. Applying the model of Rowley [95], this Miocene Δδ18Ow between Lunpola and the Siwaliks corresponds to an elevation difference of $\sim 2770 \pm 530$ m [125]. The results are consistent with recent estimates from pollen and mammal fossil studies [13,30].

By the paleoaltimeters of the oxygen and hydrology isotopes, and fossils, the northern Lhasa terrane probably began to uplift when the Bangong–Nujiang Ocean closed after the Late Cretaceous. Although the tectonic model suggested an Altiplano plateau existed during Late Cretaceous time [104,105], the paleoelevation results from the Tangrayum Co Basin, Nima, and Lunpola Basin in the northern Lhasa terrane show they began to rapid uplift from an relatively low-elevation palaeaplain of $\sim 2500$ m by $\sim 50–46$ Ma [131,132] and had obtained near-present elevations as late as Late Eocene to Late Oligocene [12,16] or even later than 18 Ma [13,30,125]. In contrast, the basins from the Gangdese Mountains [10,14,19,103] reveal sustained high-elevation mountain ranges since the Late Cretaceous. Therefore, different parts of the Lhasa terrane probably underwent different uplift histories (Fig. 3).

The early Cenozoic sedimentary rocks deposited in the residual intermountain terrestrial basins in the Heihuling area at a modern elevation of $\sim 5000$ m, north of Shuanghu Graben, record the histories of tectonic evolution, denudation, and paleoelevation of the Qiangtang terrane in the central Tibetan Plateau [133]. The Kangtuo Formation crops out mainly in the southern Qiangtang terrane and consists mainly of purple thick-bedded conglomerates. The Suonahu Formation crops out in the northern Qiangtang terrane and is dominated by purple sandstone and thick-bedded gypsum. The Suonahu Formation unconformably overlies the Kangtuo Formation and is less deformed. The upper volcanic rocks and basal granite provide limits on the depositional age of the Kangtuo and Suonahu formations [23].

Xu et al. [23] presented stable isotope results of fluvial/lacustrine carbonate cement, pedogenic carbonate, and marl from the Kangtuo and Suonahu formations deposited between $\sim 51$ and 28 Ma in the Heihuling area of the northern Qiangtang terrane. The lithofacies associations indicate that the Kangtuo Formation was deposited in alluvial fan and fluvial flood plain environments, and the Suonahu Formation was deposited in near shore lacustrine, playa-lake, and channelized fluvial environments.

Using the isotope conglomerate test of DeCelles et al. [12], the isotopic data from fluvial cement and pedogenic carbonate nodules may be interpreted to accurately represent the original isotopic composition of surface water during the period between the Eocene and Early Oligocene. The lowest δ18Oc values of $-13.1 \%$o from the Kangtuo Formation and $-11.8 \%$o from the lower Suonahu Formation are used for regional paleoaltimetry reconstructions. Based on the modern data [15], they used 10°C as the formation temperature of soil carbonate to calculate the δ18Ow values. Based on the mean paleoease surface temperature [94], they assumed that ancient MAT was 11°C i.e. 1°C higher than the modern period in this region. The reconstructed most negative δ18Ow values are corrected $-8.25 \%$ for the two formations to make the minimum estimates of the average paleoelevation of the drainage basin. Using both the empirical model found in southern Tibet [134] and a modified theoretical model [15], the paleoelevation of the northern Qiangtang terrane is reconstructed as above 5000 m by at least the Middle Oligocene (28 Ma), similar to the present elevation in this area (Fig. 3).

Couple with the paleoelevations of the Lhasa terrane, a paleogeomorphic scenario for Eocene Tibet proposes the existence of two $>4500$ m-high mountain ranges, the Gangdese mountains to the south and the Qiangtang mountains to the north, which sandwiched a low-elevation basin in the northern Lhasa terrane (Fig. 4) [13,14].

**PALEOGENE: LOW OR HIGH HOH XIL BASIN?**

The Hoh Xil Basin, with an average elevation higher than 4700 m, is situated in the northern Qiangtang
terrane, bounded by the Tanggula Mountains to the south and the Kunlun Mountains to the north. The sedimentary basement of the Hoh Xil Basin includes slate, phyllite and metasandstone of Carboniferous, Permian and Triassic origin, and the Cretaceous and Cenozoic strata include the Fenghuoshan Group, Yaxicuo Group, and Wudaoliang Formation from the bottom up [135].

Cyr et al. [11] reported that data from lacustrine limestones of the Fenghuoshan Group. Ca/Mg ratio and C isotope data are consistent with these lakes being open and not significantly affected by evaporation, suggesting that their oxygen isotopic compositions should be reflective of in-flowing surface waters. Their analysis of Fenghuoshan carbonates shows $\delta^{18}O$ values range from $-11.7$ to $-10.3\%$ (VPDB) and $\delta^{13}C$ values range between $-7.1$ and $-2.2\%$. Model results using the isotopic data from Fenghuoshan carbonates indicate that the hypsometric mean elevation of the drainage basins feeding Hoh Xil lakes was equal to or lower than 2000 m.

Currently there is no direct way to calculate the temperature of a given paleolake at the time of carbonate precipitation. Modern carbonate precipitates at water temperatures ranging between 15° and 30°C [136, 137], and the average $\delta^{18}O$ of Fenghuoshan carbonates of $-11.1\%$, calculations of the $\delta^{18}O$ of Fenghuoshan paleolake waters using the calcite-water fractionation factor of Friedman and O’Neil [100] yield an average composition of $-9.7 \pm 2.8\%$ [11]. They have also lacked reported low-elevation paleoecologic water data west-southwest of the Tibetan Plateau to compare with their Fenghuoshan carbonate data. They had to adopt a low-elevation value of $\delta^{18}O$ of $-6\%$ following Rowley et al. [17] and Currie et al. [10]. In addition, the lacustrine limestones of the Fenghuoshan Group have been questioned [95].

Recently the new U-Pb ages of the volcanic rocks embedded in the lacustrine successions of the Fenghuoshan Group [138] coupled with the magnetostatigraphic data [139] constrained deposition of the Fenghuoshan Group to Late Cretaceous to Early Eocene rather than Early Eocene to Oligocene, which made us reconsider the paleoaltimetry histories of the Hoh Xil Basin.

In another study, samples from the Miocene Wudaoliang Formation yielded higher paired carbonate $\delta^{18}O$- and lipidd13D-based paleoelevations between 3400 and 3600 m using the same moisture source and isotope–elevation relationship as the Lumphola samples [34]. Miocene paleoaltimetry estimates from the modern gradient are unrealistically high ($\sim$8000 m), perhaps due to more southerly paleoelevations and decreased moisture recycling and aridity. If a northerly moisture source fed the northern plateau but the isotope-elevation gradient was similar to the modern gradient on the southern plateau, paleoelevations are 2000–2600 m and 4000–4200 m for the Fenghuoshan Group and Wudaoliang Formation, respectively. In the study of Polissar et al. [34], the propagated uncertainties in water $\deltaD$ values and paleoelevations are large when calculated from D/H ratios of plant wax n-alkanes. These uncertainties derive from uncertainties in the isotopic analyses, water-lipid fractionation, evaporation of soil/leaf waters, and different moisture sources for precipitation.

Quade et al. [15] indicated that there is no apparent correlation between elevation and the $\delta^{18}O$ value of flowing surface waters on the Tibetan Plateau. Both surface waters and soil carbonates display a northward increase in $\delta^{18}O$ values, of about 1.5%/° north of the Himalayan crest, even though elevation increases modestly. The isotopic increase with latitude reduces the isotope-elevation gradient for water in the northernmost plateau to $-1$ to $-2%/km$. Carbonates in both soils and lakes form at higher temperatures than assumed by previous studies on the plateau. Temperature estimates from clumped-isotope ($\Delta^47$) analyses of modern soil carbonates significantly exceed mean annual air T and modeled maximum summer soil temperatures by $15.8 \pm 2.8^\circ C$ and $9.7 \pm 2.5^\circ C$, respectively. Similarly elevated temperatures best account for the $\delta^{18}O$ values observed in modern soil and lake carbonates.

As a result, basin catchment paleoelevation yields results closer to 4000 m for the Hoh Xil Basin, by accounting for its position 6.6° north of the Himalayan crest and modest changes in the assumed value for paleowater at sea-level in the moisture source region [15], in contrast to 2000 m [11] indicated by isotope data, or 1400/2600 m suggested by D/H ratios of lipid biomarkers [34]. On the other hand, paleontological evidence argues that the Hoh Xil Basin did not reach an elevation of about 3000 m until the Early Miocene.

Miao et al. [140] reported a palynological record from the Yaxicuo Group, and analyzed its implications for stratigraphic age, paleoclimate, and paleoelevation. Before 36.4 Ma, low percentages of conifers (especially Picea) indicated that the highest elevation of the basin or the surrounding mountains was not over 1500 m a.s.l., excluding any obvious MAT change (e.g. $< 1^\circ C$). During the Late Eocene, the palynological assemblages of the Yaxicuo Group are similar to those of the Qaidam, Xining, Jiuquan, Tuha, and Hetao basins, indicating similarly arid climates.
dominated by a northwestern Chinese subtropical high, and a relatively low paleoelevation lower than 2000 m.

Sun et al. [31] reported that a fossil Berberis leaf from the Early-Middle Miocene sediments of the Wudaoliang Formation in the Hoh-Xil Basin, recovered from a present-day elevation of 4600 m, and the detailed comparison with fossil and extant species of Berberis indicates that the fossil cannot be distinguished from extant B. asiatica in leaf architectural characters. The modern B. asiatica is confined to elevations of 914–2450 m. The NLR concept assumes that the fossil and its nearest living species B. asiatica occupied similar or identical niches and lived at correspondingly similar elevations. Normally such an assumption cannot be justified but extensive analysis of fossil occurrences in Eurasia spanning the Miocene suggests that the genus Berberis does not occur as an outlier and thus can be regarded as a conservative genus.

To correct for secular climate differences between the Middle Miocene and the present, Sun et al. [31] used both modeled lapse rates and enthalpy [122]. Assuming that the Middle Miocene was 2.89°C warmer than now [141] and taking the lapse rate to be 0.601°C/100 m [19], the maximum elevation at which B. asiatica could have grown in the Early Miocene was 481 m higher than present due solely to this climatic difference. Therefore, when this plant was flourishing in the Early Miocene the paleoelevation of the Hoh Xil Basin would have been no more than 1395–2931 m.

**KUNLUN MOUNTAINS: A YOUNGER MOUNTAIN RANGE**

The Kunlun Pass Basin is located in the Kunlun Pass area of the East Kunlun Mountains on the northern Tibetan Plateau. The modern elevation of the basin is about 4600–5300 m a.s.l. The basin is filled with Pliocene and Quaternary alluvium, lacustrine and glacial deposits [142].

Wang et al. [143] reported that the results of a stable isotope study of a Late Pliocene fauna discovered in the Kunlun Mountain Pass area (~4700 m a.s.l.). The δ¹³C values of enamel samples from modern herbivores from the Kunlun Pass Basin range from −14.8 to −10.6‰, with a mean of −12.0 ± 0.7‰, indicating pure C₃ diets consistent with the current dominance of C₃ vegetation in the area. In contrast, enamel samples from fossil herbivores yielded δ¹³C values of −10.2 to −5.4‰ (with a mean of −7.9 ± 1.3‰), significantly higher than those of modern herbivores in the area. The higher δ¹³C values indicate that these ancient herbivores, unlike their modern counterparts, had a variety of diets ranging from pure C₃ to mixed C₃/C₄ vegetation. The local ecosystems in the Kunlun Pass area in the Late Pliocene likely included grasslands that had small amounts of C₄ grasses. So the elevation of the Kunlun Pass in the Late Pliocene likely was lower than that of the present day.

The δ¹⁸O values of fossil mammalian bones collected from the Kunlun Pass fossil localities are −8.5 to −10.6‰, with a mean of −9.8 ± 0.6‰. Using the approach of Zanazzi et al. [8], Wang et al. [143] estimated the paleotemperatures from the δ¹⁸O values of fossil bones and the paleo-δ¹⁸O raw derived from the δ¹⁸O of fossil tooth enamel of large herbivores. The estimated MAT for the Late Pliocene is about 10 ± 8°C, which is significantly higher than the present-day MAT of −6 to −7°C in the area. It is important to note that this approach of using the δ¹⁸O of fossil bone carbonates as a paleothermometer [8] assumes that the δ¹⁸O values of their fossil mammalian bones record early diagenetic conditions near the surface environment, reflecting local temperature as well as the local δ¹⁸O raw. Thus, the reliability of the paleotemperature estimates depends on the validity of this underlying assumption.

Although reliance on the δ¹⁸O of fossil bone carbonate as a paleothermometer entails an assumption that has yet to be validated, the paleotemperature estimates of Wang et al. [143] are broadly compatible with those from aquatic plants, ostracods and mollusk shells (~10–17°C). Assuming that (i) the temperature gradient of 0.5°C/100 m determined from the present-day conditions of the Kunlun Pass Basins applied to the past and (ii) a temperature drop of 3°C in the area was due to global cooling since the Pliocene [144], the estimated temperature change in the basin would correspond to an elevation change of ~2700 ± 1600 m since the Late Pliocene [143]. This would imply that the elevation of the Kunlun Pass Basin in the Late Pliocene was ~2011 ± 1600 m a.s.l., much lower than its present-day elevation.

Wang and Chang [115] described the cyprinid fossils from the Pliocene lower member of Qiangtang Formation of the Kunlun Pass Basin, collected at a locality 4769 m a.s.l. The materials consist of numerous disarticulated and incomplete bones as well as thousands of pharyngeal teeth, fin rays, and vertebrae. Some fossils were referred to the genus Gymnocypris, lineage Schizothoracini, family Cyprinidae, of the highly specialized grade. Two species are so far unequivocally assigned to Gymnocypris, G. przewalskii and G. eckloni, and they inhabit Qinghai Lake and the waters both north (the Golmud River) and south (upper reach of the Yellow River) of the East...
Kunlun Mountain, respectively. *Gymnocypris ekkoni* can reach the elevation of 4200 m, and the habitat of *G. przewalskii*, the Qinghai Lake, is at a relatively low elevation of 3200 m [108].

Since representatives of the genus appeared abundantly in the area of the Kunlun Pass Basin during the Pliocene, it is natural to assume that the elevation of the watershed then was probably somewhere between 3200 and 4200 m, lower than the present height of the locality. In this case, it is not irrational to propose that the Kunlun Pass Basin has risen approximately 1000 m since the Pliocene [115]. This deduction is probably less ambiguous than the suggestion above of 2700 ± 1600 m [143]. It is also different from the estimation of [145], i.e. the elevation of the Kunlun Pass was 1000–1500 m during the period of 7.0–1.1 Ma, based on studies of fossil ostracods [146,147] and pollens [148].

Wang and Chang [149] described the nemacheilid fish fossils from the Kunlun Pass Basin. These bones are very similar to their counterparts in some species of a modern nemacheilid genus, *Triplophysa*, which is widely distributed on the Tibetan Plateau. The nemacheilid fossils are much more abundant than the remains of schizothoracines embedded in the same horizon at the same locality [115]. This would imply that the number of individuals of *Triplophysa* was much higher than that of schizothoracines when they were alive in the area. In the modern ichthyofauna of the Tibetan Plateau, *Triplophysa* prevails over schizothoracines in the number of individuals in the high elevations and small water bodies. Some *Triplophysa* species have a greater adaptability to the colder environment with shallow waters than schizothoracines, and thus have a wider distribution to higher elevation (e.g. *T. microps*, up to 5600 m a.s.l.) than the latter [150].

Based on the fossil dominance of *Triplophysa* over schizothoracines and their taphonomical conditions, it appears that the water system at the Kunlun Pass area during the Pliocene might not be extensive lakes or large rivers with broad valleys. There might have been a few mountainous, relatively torrential rivers with many small, shallow streams connecting the water systems from the north and south of the East Kunlun Mountain. The environment of the Kunlun Pass Basin area during the Pliocene would have been very harsh, and the elevation of the area might already have been higher than previously suggested based on schizothoracines [115]. As a result, the uplift of the area may have been less than 1000 m since the Pliocene [149].

**SUMMARY**

In conclusion, the timing of uplift of the Tibetan Plateau remains controversial, especially in a particular basin or area in the plateau. To view the Tibetan Plateau as a whole, requires a different, complex interpretation. The mammalian fossils discovered from both sides of the Tibetan Plateau also imply a high plateau since the late Early Miocene (Fig. 5). During the Oligocene, giant rhinos lived in northwestern China, north of the Tibetan Plateau, while they were also distributed in the Indo-Pakistan subcontinent to the south of the Tibetan Plateau, which indicates that the elevation of the Tibetan Plateau was not too high to prevent exchanges of large mammals; giant rhinos, the rhinocerotid *Aprotodon*, and chalicotheres still dispersed north and south of ‘Tibetan Plateau’ [27]. In contrast, during the Middle Miocene, the shovel-tusked elephant *Platybelodon* was found from many localities north of the Tibetan Plateau, while its trace was absent in the Siwaliks of the subcontinent, which implies that the Tibetan Plateau had uplifted high enough to obstruct the exchange of mammals in the Middle Miocene [74,151].

On the other hand, most paleoelavations reconstructed based on $\delta^{18}$O values of carbonates in the Tibetan Plateau are close to modern elevations at least during the Late Oligocene to Middle Miocene [4,10,12,14–17,23,48,95,103,152]. However, the isotopic paleoelavimeter requires many assumptions for a series of uncertain parameters and conditions, such as age control, diagenesis, sample type, evaporation effects, assumed paleotemperatures, and climate change effects, which may lead to erroneous interpretations of paleoelavation. From the perspective of oxygen isotopes, for example, the Tibetan Plateau can be divided into two regions: the southern region where oxygen isotope lapse rates are steep and the northern region where lapse rates are half those of the south [153,154]. For each Cenozoic basin, different researchers have different views about which lapse rate should be adopted for the Paleogene or for the Neogene.

The paleoelavimetry reconstructions and history are crucial to interpret the geodynamic evolution and to understand the climatic changes in Asia, but precise results have a long way to go for researchers because several controversial issues remain unresolved. In the future work, the applications of multidisciplinary comprehensive methods and the cross-checks on results are important to achieve more reliable estimates for the paleoelevations of the Tibetan Plateau.
Figure 5. South–north elevation profiles of the Tibetan Plateau from the Oligocene to the Present based on paleontological evidence, showing distribution of the Oligocene giant rhino and the Miocene shovel-tusked elephant.

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