Vol 439|9 February 2006|doi:10.1038/nature04506

Palaeo-altimetry of the late Eocene to Miocene Lunpola basin, central Tibet

David B. Rowley[1] & Brian S. Currie[2]

The elevation history of the Tibetan plateau provides direct insight into the tectonic processes associated with continent–continent collisions. Here we present oxygen-isotope-based estimates of the palaeo-altimetry of late Eocene and younger deposits of the Lunpola basin in the centre of the plateau, which indicate that the surface of Tibet has been at an elevation of more than 4 kilometres for at least the past 35 million years. We conclude that crustal, but not mantle, thickening models, combined with plate-kinematic solutions of India–Asia convergence, are compatible with palaeo-elevation estimates across the Tibetan plateau.

The elevation history of the Tibetan plateau is one of the few direct ways of discriminating among models for the modes of strain accommodation in the India–Asia continent–continent collision system. Data constraining the elevation history of the Himalaya and Tibet can have a critical role in determining the degree to which various mechanisms—including crustal subduction[1], crustal thickening[2,3], crust and mantle-lithosphere thickening with subsequent convective removal of the mantle lithosphere[4–6], as well as escape[7,8] and extrusion[9]—have contributed to the high elevation and distribution of strain across the region. In this Article, we present new data from the Lunpola basin in central Tibet (Fig. 1) that constrain an ~40 Myr elevation history of this part of the India–Asia collisional system using oxygen-isotope palaeo-altimetry of Eocene–Oligocene and Miocene palaeosol carbonates and lacustrine limestones. We summarize existing palaeo-altimetric estimates in comparison with retrodictions (predictions about the past) made within the context of India–Asia convergence history since the onset of collision.

Regional setting of Lunpola basin

The Lunpola basin is primarily a Tertiary sedimentary basin situated in the central part of the Tibetan plateau at about 32° N, 89.75° E (Fig. 1). The basin is situated approximately half way between the Qaidam basin to the north and the Himalaya to the south. The basin axis strikes essentially east–west and has been variably deformed along basin-parallel thrust faults and associated folds. Erosion through these structures has exposed much of the succession at the surface. The Lunpola basin contains a thick stratigraphic succession that extends from the Palaeocene to the Pliocene[10]. The basin has been explored geophysically and through a series of wells[11] that has established the stratigraphic context and correlation of the sediments within the basin[10,11].

The Cenozoic strata of the Lunpola basin are more than 4,000 m thick, and consist of two primary stratigraphic units: the Palaeocene–Oligocene Niubao Formation, and the Miocene–Pliocene Dingqing (also referred to as the Dingqinghu) Formation. The age of these formations is based primarily on fossil ostracode and palynological assemblages[10,12–14]. As part of our investigation, we sampled lacustrine and palaeosol carbonates from the middle and upper Niubao Formation, respectively, as well as lacustrine marls and limestones from the middle Dingqing Formation. In terms of age, the middle Niubao Formation is classified as Eocene and the upper member is late Eocene to Oligocene, while the middle Dingqing Formation is Miocene in age[10,12–14].

Our sampling in the Niubao Formation was conducted in outcrops along the northern margin of the basin (Fig. 1 inset, location B). Here about 2.2 km of the middle and upper Niubao Formation are exposed in the north limb of a fault-propagation anticline in the hanging wall of the northern basin-margin thrust. At this location, the middle Niubao Formation consists of lacustrine mudstones and dolomitic marls, with minor beds of sandstone, algal dolostone, kerogenous shale, marl and micritic limestone. Because of our desire to make estimates of palaeometeoric precipitation based on oxygen isotopic composition of carbonate, we limited our sampling to the calcareous marls and micritic limestones preserved in the lower and middle parts of the middle Niubao Formation. At this location, the upper Niubao Formation consists of >400 m of fluvial/alluvial mudstone, sandstone and conglomerate that contains abundant intercalated calcic palaeosols near the top of the section. Pedogenic carbonate nodules from this interval were extensively sampled as part of this study.

We also sampled micritic lacustrine carbonates from the middle member of the Dingqing Formation in outcrops in the south-central parts of the basin (Fig. 1 inset, location A). The studied interval consists of mudstone, marl, micritic and oolitic limestone, sandstone and conglomerate. Some sandstones and siltstones contain symmetrically-rippled surfaces, as well as fossilized fish bones. Finer-grained lithologies were in places well laminated and contained fossil ostracodes and insects. Because of low structural dips and subdued topography, only 19 m of the unit is exposed at this location.

Isotope-based palaeo-altimetry

Rowley *et al.*[15] described the physical and thermodynamic basis of a model of oxygen and hydrogen isotopes in precipitation in low latitude (<35°) orographic systems to estimate (palaeo-)elevations and potentially (palaeo-)hypsometries of hydrologic systems. Ref. 15 presents a model of the relationship between elevation and $\Delta(\delta^{18}O_p)$, where $\Delta(\delta^{18}O_p)$ is the difference in isotopic composition of precipitation between that at sea level and that at elevation. Rowley *et al.*[15] used data from the oldest intermontane basins within the Himalaya to demonstrate that the Himalaya have had their current stature for

---

[1]Department of the Geophysical Sciences, 5734 S. Ellis Avenue, The University of Chicago, Chicago, Illinois 60637, USA. [2]Department of Geology, 114 Shideler Hall, Miami University, Oxford, Ohio 45056, USA.

at least the last 8 to 10 Myr. Garzione *et al.*[16,17], using an empirical fit to surface water data, arrived at a similar conclusion.

Modern stream waters from Amdo to Lunpola

Comparison of the oxygen isotopic compositions of modern waters in southern Tibet with the model demonstrate that the modern system accords well with predictions[15,18]. Small streams were sampled along the Amdo-Dongqiao-Lunpola road and within the Lunpola basin (Fig. 1). Elevations of the sampling sites range from 4,567 m to 4,718 m, based on GPS estimates. Maximum drainage basin heights are only a few hundred metres to a kilometre or so higher than the sample elevation, and hence there should not be a significant hypsometric effect in these samples. We compute predicted altitudes and palaeo-altitudes using equation (1) from ref. 18 (Fig. 2). Note that where samples are not directly measuring precipitation, as in surface and ground waters, we replace $\Delta(\delta^{18}O_p)$ with $\Delta(\delta^{18}O_w)$ to signify this, and to remind ourselves that $\Delta(\delta^{18}O_w)$ includes hypsometric effects of integrating precipitation in the drainage basin above the sample elevation. These Tibetan streams range from relatively unfractionated, with $\Delta(\delta^{18}O_w)$ of −3.2‰, to highly fractionated, ranging up to −11.0‰ (Fig. 2). The majority of predicted elevations of these streams (Fig. 2) overlap within uncertainty with model expectations. It is important to recognize that modern waters do not plot above the model correlation line which would indicate fractionation by some combination of Himalayan orographic lifting and the effects of 'continentality'[19,20]. In fact, other factors, not yet completely understood, reduce the altitude effect significantly for a number of these and other sample streams reported in the literature from central and northern Tibet[21]. If similar effects operated in the past, the estimated $\Delta(\delta^{18}O_w)$ would result in an underestimate of the palaeo-elevation in this region. Given our current data, we have yet to encounter waters that are more depleted than predicted by the model[15]. We take this as a positive bias, as it leads to the expectation that the isotopes are much more likely to underestimate, rather than overestimate, (palaeo-)elevations.

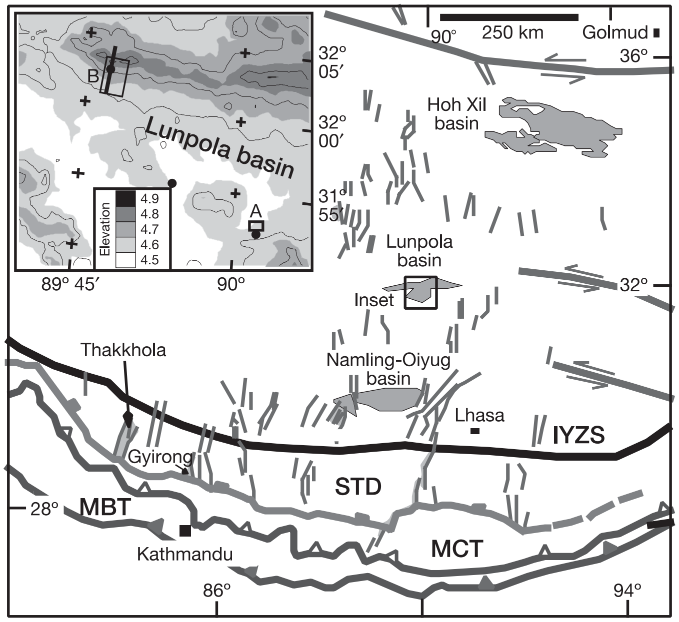

Figure 1 | Generalized tectonic map of the Himalaya–Tibet region, and details of the Lunpola basin. In the main figure are shown basins within the plateau for which palaeo-elevation estimates discussed in the text are derived. The location of the inset is shown as a box. Inset, map showing details of the Lunpola basin. Topographic contours are derived from Hydro1K Asian data; key shows elevation in km above sea level. Boxes labelled A and B are sites from which samples for palaeo-elevation measurements were collected for the Dingqing and Niubao, respectively. Heavy line is the location of the middle and upper Niubao section. Black circles are sites of collection of modern streams. Abbreviations: MBT, Main Boundary thrust; MCT, Main Central thrust; STD, South Tibetan detachment; IYZS, Indus-Yarlung Tsangpo suture.

Palaeo-elevation estimates of the Lunpola region

Palaeo-altitude estimates depend on $\Delta(\delta^{18}O_w)$. This requires the identification of low altitude 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 we use modern observations to place bounds on likely compositions of low latitude and low altitude precipitation, and compare these with general circulation model-based estimates incorporating isotopes. In the modern world, there is little change in isotopic composition of low latitude, near sea level isotopic compositions of precipitation. Values of $\delta^{18}O_p$ of about -3.6 ± 1.6‰ are found for average low latitude ($< \pm 35°$) and low elevation stations (< 100 m), based on data released by the IAEA Global Network of Isotopes in Precipitation for years through to 2001[22]. Given these estimates, we adopt a value of $\delta^{18}O_p$ of -6 ± 1.6‰ for the low elevation precipitation, comparable to the modern precipitation in New Delhi and more depleted than the average low latitude, low elevation, coastal value observed today. By adopting a more depleted low elevation value, we decrease our estimated palaeo-elevations.

The Dingqing Formation represents the youngest pre-Quaternary strata in the Lunpola basin. Our sampling focused on thin-bedded marls and fine grained limestones of the Miocene middle Dingqing Formation. Oxygen isotopic compositions of the lacustrine carbonate ($\delta^{18}O_c$) range from −1.3‰ to −14.6‰ (with respect to the Pee Dee belemnite (PDB) standard). It is important to note that the most widely ranging values in this section come from samples of thin (3–5 cm thick) marls that are only 50 cm apart. Figure 3, main panel, plots $\Delta(\delta^{18}O_w)$ versus predicted palaeo-altitudes of these Miocene samples. For each sample, we plot the mean $\Delta(\delta^{18}O_w)$ and the $\pm 2\sigma$ uncertainty in $\Delta(\delta^{18}O_w)$. This reported uncertainty reflects most

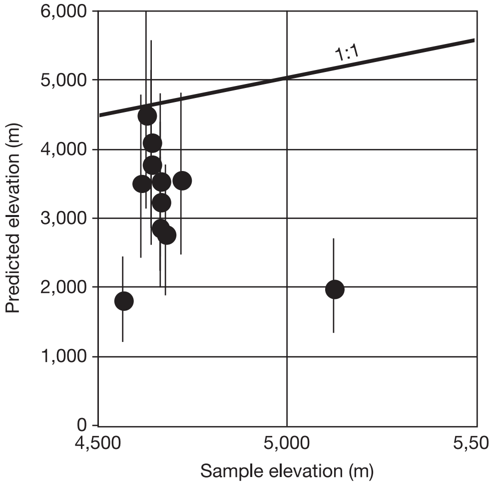

Figure 2 | Sample elevations versus predicted elevations of modern stream waters in central Tibet. Sample elevations are based on GPS estimates of elevations at the sample site. Predicted elevations (h) are based on measured oxygen isotopic composition differenced with the weighted mean annual composition of precipitation in New Delhi ($\delta^{18}O_p = -5.8$‰) and:

$$h = -6.14\times10^{-3}\Delta(\delta^{18}O_p)^4 - 0.6765\Delta(\delta^{18}O_p)^3 - 28.623\Delta(\delta^{18}O_p)^2 - 650.66\Delta(\delta^{18}O_p) \quad (1)$$

The line of 1:1 correlation is included. Error bars are model-related 2σ uncertainties of elevation based on $\Delta(\delta^{18}O_w)$ (ref. 15).

importantly uncertainty in the temperature at which carbonate and water equilibrated, and the corresponding uncertainty in the estimate 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[15]. Besides orographic ascent, probably the largest additional controller on $\Delta(\delta^{18}O_w)$ is evaporation. Evaporation preferentially removes ^{16}O, resulting in an enrichment of $\delta^{18}O_w$ in the remaining water and a reduction in $\Delta(\delta^{18}O_w)$. The best estimate of the Miocene palaeo-altitude of the Lunpola basin is thus provided by the most depleted $\Delta(\delta^{18}O_w)$ values, as these are probably the least affected by evaporative re-enrichment or other atmospheric processes that reduce the depletion relative to that predicted by the model. The most depleted samples with a $\Delta(\delta^{18}O_w)$ of -10.0 ± 2.0‰ yield an estimated palaeo-elevation of 4,260 m +475/−575 m.

Upper Niubao Formation samples are derived from micritic $CaCO_3$ nodules from stacked argillic calcisols. Extensive sampling of multiple carbonate nodules from 12 different horizons at one locality (no. 305), as well as additional samples from overlying palaeosol carbonate nodule-bearing horizons, provide 60 separate analyses of $\delta^{18}O_c$ and $\delta^{13}C_c$. The average $\delta^{13}C_c$ of all of the samples is -7.6 ± 0.7‰ (2σ), consistent with development associated with C_3 vegetation. $\delta^{18}O_c$ averages -17.2 ± 1.2‰ (2σ). This consistency means that it does not matter whether the individual samples are grouped by locality or by horizon; as shown in Fig. 3, the results are similar. The mean $\Delta(\delta^{18}O_w)$ for these samples is -12.6 ± 1.2‰ (2σ), corresponding with a predicted mean elevation of the basin of about 4,850 m +380/−460 m. Model uncertainty of elevation for a $\Delta(\delta^{18}O_w)$ of −12.6‰ is 4,850 m +1,630/−1,435 m (2σ) (ref. 15).

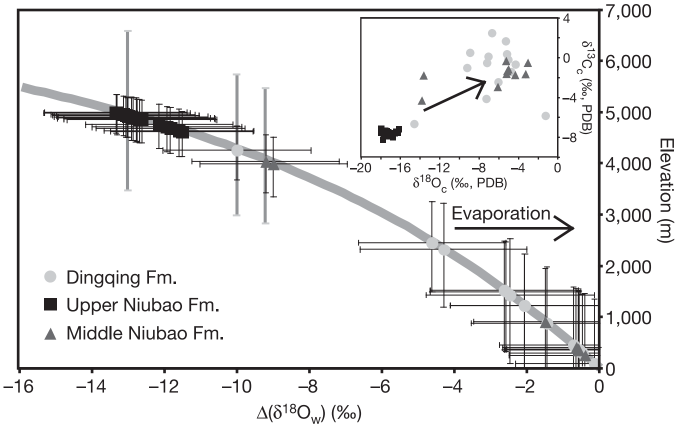

Figure 3 | $\Delta(\delta^{18}O_w)$ versus predicted elevations derived from Lunpola carbonates. Main panel, $\Delta(\delta^{18}O_w)$ versus predicted elevations of carbonate-derived estimates of local surface and soil waters from the Lunpola basin in central Tibet. Deriving $\delta^{18}O_w$ from $\delta^{18}O_c$ assumes that carbonate and water equilibrated at 10 ± 10 °C, corresponding to $\alpha_{CaCO_3-H_2O}$ of about 1.0323 +0.0027/−0.0024 for colder and warmer temperatures, respectively, based on the temperature dependence of the fractionation factor[32], and we incorporate this uncertainty in our estimates of $\Delta(\delta^{18}O_w)$. The arrow emphasizes that evaporation reduces $\Delta(\delta^{18}O_w)$, which results in a lower estimate of predicted elevation. The model curve from Rowley *et al.*[15] (based on equation (1)) is shown for reference. Two sets of uncertainties are plotted with each data point. Black, horizontal uncertainties combine uncertainties in $\delta^{18}O_c$, $\Delta(\delta^{18}O_w)$ and in the temperature of fractionation of carbonate and water. These uncertainties give rise to an uncertainty (1σ) in the estimate of the mean elevation of the sample, shown as the black vertical bars. Lighter grey vertical bars, shown only for three samples, reflect 2σ uncertainties associated with starting relative humidity and temperature of the low elevation moisture source (see ref. 15 for a more complete discussion). Inset, plot of $\delta^{13}C_c$ versus $\delta^{18}O_c$ of data from each of the sample horizons. The arrow shows the expected variation associated with evaporation. Symbols are the same for each plot.

Stratigraphically beneath the upper Niubao Formation, palaeosol carbonates are a sequence of variably dolomitized and silicified thin-bedded lacustrine carbonates of which we focused our sampling on limestones and marls. $\delta^{13}C_c$ ranges from −4.4‰ to −0.3‰, while $\delta^{18}O_c$ ranges from −3.1‰ to −13.8‰ (PDB) (Fig. 3 inset). As with the Dingqing lacustrine carbonates, there is a substantial range in estimates of both $\Delta(\delta^{18}O_w)$ and predicted palaeo-elevations from these carbonates. The most negative samples, with $\Delta(\delta^{18}O_w)$ of about -9.2 ± 2.0‰, give the best estimate of the palaeo-elevation for this sequence of about 4,050 m +510/−620 m. Model uncertainties corresponding with a $\Delta(\delta^{18}O_w)$ of about −9.2‰ are 4,050 m +1,420/−1,220 m (ref. 15).

Oxygen isotopic data from the late Eocene–Oligocene Niubao Formation and the Miocene Dingqing Formation yield remarkably consistent results. The most negative, and hence least potentially evaporatively enriched, samples yield estimated palaeo-altitudes for the Lunpola basin in excess of 4,000 m, and all overlap within uncertainty with the basin's current mean elevation of about 4,600 m. The upper Niubao Formation palaeosol carbonate nodules yield the most consistent results, suggesting a highly elevated basin within which C_3 vegetation dominated the carbon isotopic budget. Both older and younger lacustrine-dominated sequences appear to be more variably affected by evaporation, but nonetheless the most depleted oxygen isotopic compositions also have the most negative carbon isotopic compositions, consistent with limited evaporative or secondary alteration that suggest comparably high elevations. The most reasonable interpretation is that this part of the Tibetan plateau was already elevated to its current altitude by the time of deposition of the upper Niubao in late Eocene to early Oligocene times. Central to the above interpretations is the assumption that the stable-isotopic composition of Lunpola basin lacustrine and palaeosol carbonates accurately record the composition of meteoric waters at the time of deposition. Diagenetic alteration associated with high-temperature recrystallization or younger more depleted meteoric waters may result in a reduction in $\delta^{18}O$ compositions that may lead to erroneous interpretations of palaeo-elevation[23].

Several lines of evidence lead us to believe that overall digenetic alteration of our samples has been minimal. First, all of our samples were derived from micritic carbonates that display little evidence of recrystallization. While the effects of diagenesis on carbonates can be subtle, our samples do not display the pervasive microcrystallization reported from the mid-Cenozoic basins of eastern Tibet[23]. In addition, as a whole, our samples display a positive covariance of $\delta^{18}O$ and $\delta^{13}C$ values (Fig. 3 inset). We interpret this trend to reflect evaporative enrichment of meteoric waters in Lunpola lacustrine systems throughout deposition. Similar trends from other isotopic studies have been interpreted to represent lacustrine carbonates that have undergone minimal diagenetic alteration[23]. The variability in isotopic composition does not display any vertical stratigraphic trends that might be attributed to burial diagenesis. In fact, the greatest observed $\delta^{18}O$ variations in the both the middle Niubao and Dingqing formations come from samples that are in close stratigraphic proximity (15 m and 50 cm, respectively). As such, it is unlikely that any diagenetic or more recent secondary alteration has effected the isotopic compositions of these carbonates.

Discussion

The palaeo-elevation estimates derived from the Lunpola basin are important for several reasons. First, these are currently the oldest data indicating that parts of the plateau were highly elevated (>4 km) by the late Eocene to early Oligocene. The consistency of the data from this basin over time supports models in which crustal rheology limits elevation and that once elevated the plateau surface remains at essentially the same elevation over time[4,9,24]. Second, these data are consistent with the conclusion[18,25] that models invoking significant mantle involvement through initial mantle-lithosphere thickening and subsequent convective destabilization and removal of this mantle lithospheric root[4–6] are not supported by any palaeo-elevation data available from the plateau.

An important question remains: are the estimates of the palaeo-elevation history of the Lunpola basin expected on the basis of current understanding of the controls on orogenic plateau development? This can only be addressed if explicit retrodictions of the growth of the plateau can be formulated. Figure 4 presents an estimate of the retrodicted increase in width of the plateau as a function of time. Our estimate derives from the model of Beaumont *et al.*[3], as it makes the most explicit retrodictions currently available. The modelled[3] width of the entire orogen in excess of 4 km elevation increases by about 40% of the cumulative convergence for the interval from ~54 Myr to 30 Myr after initiation of collision; this phase of the model includes erosion at the front, and a crust–mantle density contrast of 500 kg m^{-3}. The entire width of the orogen is not preserved as a function of time owing to erosion, primarily along the Himalaya. We therefore measure width relative to the Indus-Yarlung Tsangpo suture, as it is more readily defined as a function of time and its position is tracked in the model[3]. Approximately 80% of the crustal mass of the plateau lies north of the suture and hence we further reduce the scaling of retrodicted plateau width by this factor, resulting in an overall scaling of plateau width north of the suture to 32% of the cumulative convergence. We use the independently determined India–Asia convergence history since the initiation of collision to scale the predicted width of the plateau in Fig. 4. Initiation of the India–Asia collision occurred at 50.6 ± 0.2 Myr ago[26] in the vicinity of Mount Everest. No data tightly constrain the age of initiation of collision farther east[27]. Cumulative collision-related convergence varies as a function of position along the suture, but at 90° E the total is about 3,500 km (Fig. 4); it increases to about 3,700 km in the east and decreases to about 2,650 km in the west. Scaling by 40% predicts a current plateau width of 1,400 km and a width north of the suture of about 1,100 km, both of which accord well with observation. Between the time of initiation of the collision and magnetic reversal chron 21 (at ~46.2 Myr ago; ref. 28), the India–Asia convergence rate was approximately twice the subsequent rate, and hence the scaling requires an early, more rapid increase in width of the plateau, followed by a more steady northward growth at an average of about 21 km Myr^{-1} (Fig. 4) since about 46.2 Myr ago.

The ages at which various basins situated north of the suture are retrodicted to have exceeded 4 km elevation are also plotted in Fig. 4. Palaeo-elevation estimates are available for each of these basins, but unfortunately only the data from the Lunpola and Hoh Xil basins are derived from sufficiently old sediments to provide much of a test of the retrodictions. The Lunpola basin is retrodicted to have achieved an elevation in excess of 4 km by ~39 Myr ago, and the middle and upper Niubao data that date from the late Eocene to Oligocene (~35 ± 5 Myr ago) are compatible with this estimate (Fig. 4). Cyr *et al.*[29] report data from lacustrine limestones of the Fenghuoshan Group[29] of late Eocene age (~39 Myr ago) in the Hoh Xil basin (Fig. 1). Ca/Mg 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. Palaeo-elevation estimates from Fenghuoshan lacustrine carbonates suggest that this part of the plateau was at elevations below 2 km at 39 Myr ago, again consistent with the model which retrodicts elevation above 4 km of the Hoh Xil basin region at about 25 Myr ago. Unfortunately, there are as yet no younger materials that have yielded palaeo-elevation estimates available from this part of the plateau. The Qaidam basin is retrodicted to have been elevated above 4 km at about 13 Myr ago. The Qaidam basin is currently at an average elevation below 3 km, but shortening jumped across the Qaidam basin to the Qilian Shan at about this time[8], isolating the Qaidam basin within the broader Tibetan orogenic system. Isotopic data from the Tarim and Qaidam basins are broadly consistent with changing orography and moisture source input at this time[30]. A palaeo-enthalpy-based estimate[25] and an oxygen-isotope-based estimate[18] agree with each other, and place the Oiyug basin (situated just north of the suture) above 4 km elevation at 15 Myr ago. The model retrodicts the uplift of this region shortly after collision and hence these data, although consistent with expectation, do not help to constrain the model.

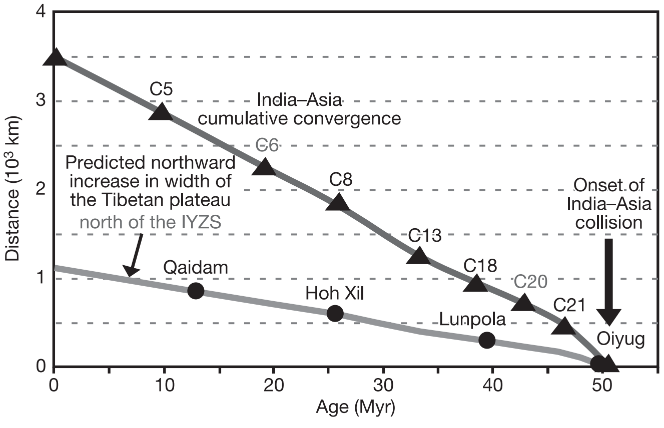

Figure 4 | Predicted elevation history of the Tibetan plateau north of the Indus-Yarlung Tsangpo suture (IYZS) at the longitude of 90° E, and the India–Asia convergence. Plateau width is scaled on the basis of thermo-mechanical modelling to the convergence history of the India–Asia collision. India–Asia convergence history is constrained at each of the magnetic reversals (filled triangles) represented by C5 through to C22 (ref. 28). Predicted ages of uplift of various basins above 4 km height across the plateau are shown by the filled circles. Locations of the basins are shown in Fig. 1.

Currently available data are consistent with a simple model that scales India–Asia convergence to increase in plateau width as a function of time. It is extremely encouraging that palaeo-elevation data from the Himalaya and Tibet are beginning to be able to discriminate among various models that have been proposed in the past. A significant finding in this regard is the absence of evidence for mantle lithosphere thickening and convective removal as a contributor to the elevation history of the plateau. At present, the predictions based on thermal-mechanical modelling[3] are consistent with the existing data. Given our current data and understanding, estimates of plateau growth should be based on progressive northward growth of the plateau at between 4 km and 5 km elevation at a rate of about 20 km Myr^{-1} after about 46.2 Myr ago. Regions farther to the east involve additional contributors to the strain accommodation history that make scaling of convergence to plateau width less straightforward. Regions that lie east of the convergence trajectory of India relative to Asia[31] would not be expected to be elevated according to the simple model presented here. Eastward export of crustal mass, primarily through sub-crustal flow[9], contributes a sink that does not exist in the centre of the plateau in the vicinity of Lunpola. Additional contributions from crustal and lithospheric escape may also be significant. Together, these suggest that increase in width in these more eastern regions may not be so simple. A final comment is to emphasize that there is no evidence supporting plateau-wide uplift at 8–10 Myr ago and hence any suggested correlates associated with previous model-derived[6] suggestions now need to be re-examined.

Overview

Oxygen-isotope-based estimates of palaeo-altitude from the late Eocene–Oligocene Niubao Formation and Miocene middle Dingqing Formation of central Tibet indicate that the central Tibetan plateau has been characterized by elevations in excess of 4 km since 35 ± 5 Myr ago. Elevations in excess of 4 km are compatible with expectations if recent thermo-mechanical modelling[3] are combined with estimates of the India–Asia age of collision and convergence to provide first-order retrodictions of the elevation history of the Tibetan plateau. In the vicinity of the Lunpola basin, these retrodictions would imply that an elevated plateau existed by at least 39 Myr ago, and the data reported here are compatible with this. Data from the Lunpola basin further imply, again consistent with modelling, that once elevated the plateau's mean elevation did not vary within uncertainty. Finally, data from the Lunpola basin are consistent with observations derived from the Oiyug basin to the south[18,25] that there is no evidence that the plateau underwent a plateau-wide uplift in the late Miocene (~8–10 Myr ago) associated with convective destabilization of a thickened lithospheric-mantle root beneath the plateau. Rather, all existing data are consistent with models in which the mantle lithosphere is simply subducted into the mantle.

Received 28 August; accepted 29 November 2005.

1. Argand, E. La tectonique de l'Asie. In *Proc. 13th Int. Geological Congress* Vol. 1, Part 5, 171–372 (Vaillant-Carmanne, Liège, 1924).
2. Dewey, J. F. & Burke, K. C. A. Tibetan, Variscan, and Precambrian basement reactivation: products of continental collision. *J. Geol.* **81,** 683–692 (1973).
3. Beaumont, C., Jamieson, R. A., Nguyen, M. H. & Medvedev, S. Crustal channel flows: 1. Numerical models with applications to the tectonics of the Himalayan–Tibetan orogen. *J. Geophys. Res.* **109,** B06406, doi:10:1029/2003JB002809 (2004).
4. England, P. & Housemann, G. Finite strain calculations of continental deformation. II: Application to the India–Asia plate collision. *J. Geophys. Res.* **91,** 3664–3676 (1986).
5. Houseman, G. & England, P. in *The Tectonic Evolution of Asia* (eds Yin, A. & Harrison, T. M.) 3–17 (Cambridge Univ. Press, Cambridge, 1996).
6. Molnar, P., England, P. & Martinod, J. Mantle dynamics, uplift of the Tibetan plateau, and the Indian Monsoon. *Rev. Geophys.* **31,** 357–396 (1993).
7. Tapponnier, P., Peltzer, G., LeDain, A. Y., Armijo, R. & Cobbold, P. Propagating extrusion tectonics in Asia: new insights from simple experiments with plasticine. *Geology* **10,** 611–616 (1982).
8. Tapponnier, P. *et al.* Oblique stepwise rise and growth of the Tibet plateau. *Science* **294,** 1671–1677 (2001).
9. Royden, L. H. *et al.* Surface deformation and lower crustal flow in eastern Tibet. *Science* **276,** 788–790 (1997).
10. Bureau of Geology and Mineral Resources Xizang Autonomous Region. *Regional Geology of Xizang (Tibet) Autonomous Region* (Geological Publishing House, Beijing, 1993).
11. Zhai, G. *et al.* (eds) Oil and gas fields in Qinghai and Xizang. In *Petroleum Geology of China* Vol. 14 (Petroleum Industry Press, Beijing, 1990).
12. Xia, J. B. Cenozoic of Baingoin and its borders, Xizang (Tibet) [Chinese with English summary]. *Contrib. Geol. Qinghai-Xizang (Tibet) Plateau* **3,** 243–254 (1983).
13. Xu, Z. Y. The Tertiary and its petroleum potential in the Lunpola Basin, Tibet. *Oil Gas Geol.* **1,** 153–158 (1980).
14. Xu, Z. Y., Zhao, J. P. & Wu, Z. L. On the Tertiary continental basins and their petroleum potential in Qinghai-Xizang (Tibet) Plateau with Lunpola Basin as example [Chinese with English summary]. *Contrib. Geol. Qinghai-Xizang (Tibet) Plateau* **17,** 391–399 (1985).
15. Rowley, D. B., Pierrehumbert, R. T. & Currie, B. S. A new approach to stable isotope-based paleoaltimetry: Implications for paleoaltimetry and paleohypsometry of the High Himalaya since the Late Miocene. *Earth Planet. Sci. Lett.* **188,** 253–268 (2001).
16. Garzione, C. N., Dettman, D. L., Quade, J., DeCelles, P. G. & Butler, R. F. High times on the Tibetan Plateau: Paleoelevation of the Thakkhola graben, Nepal. *Geology* **28,** 339–342 (2000).
17. Garzione, C. N., Quade, J., DeCelles, P. G. & English, N. B. Predicting paleoelevation of Tibet and the Himalaya from δ^{18}O vs. altitude gradients in meteoric water across the Nepal Himalaya. *Earth Planet. Sci. Lett.* **183,** 215–229 (2000).
18. Currie, B. S., Rowley, D. B. & Tabor, N. J. Middle Miocene paleoaltimetry of southern Tibet: Implications for the role of mantle thickening and delamination in the Himalayan orogen. *Geology* **33,** 181–184 (2005).
19. Dansgaard, W. Stable isotopes in precipitation. *Tellus* **16,** 436–468 (1964).
20. Rozanski, K., Araguas-Araguas, L. & Gonfianti, R. in *Climate Change in Continental Isotopic Records* 1–36 (American Geophysical Union, Washington, 1993).
21. Tian, L., Masson-Delmotte, V., Stievenard, M., Yao, T. & Jouzel, J. Tibetan Plateau summer monsoon northward extent revealed by measurements of water stable isotopes. *J. Geophys. Res.* **106,** 28081–28088 (2001).
22. IAEA. ⟨http://isohis.iaea.org/userupdate/GNIP2001yearly.xls⟩ (2001).
23. Garzione, C. N., Dettman, D. L. & Horton, B. K. Carbonate oxygen isotope paleoaltimetry: Evaluating the effect of diagenesis on paleoelevation estimates for the Tibetan plateau. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **212,** 119–140 (2004).
24. Beaumont, C., Fullsack, P. & Hamilton, J. in *Thrust Tectonics* (ed. McClay, K. R.) 1–18 (Chapman & Hall, London, 1992).
25. Spicer, R. A. *et al.* Constant elevation of southern Tibet over the past 15 million years. *Nature* **421,** 622–624 (2003).
26. Zhu, B., Kidd, W. S. F., Rowley, D. B., Currie, B. S. & Shafique, N. Age of initiation of the India–Asia collision in the East-Central Himalaya. *J. Geol.* **113,** 265–285 (2005).
27. Rowley, D. B. Age of initiation of collision between India and Asia: A review of stratigraphic data. *Earth Planet. Sci. Lett.* **145,** 1–13 (1996).
28. Berggren, W. A., Kent, D. V., Swisher, C. C. III & Aubry, M. P. in *Time Scales and Global Stratigraphic Correlation* (eds Berggren, W. A., Kent, D. V., Aubry, M. P. & Hardenbohl, J.) 129–218 (Society of Economic Paleontologists and Mineralogists, Tulsa, Oklahoma, 1995).
29. Cyr, A., Currie, B. S. & Rowley, D. B. Geochemical and stable isotopic evaluation of Fenghuoshan Group lacustrine carbonates, north-central Tibet: Implications for the paleoaltimetry of Late Eocene Tibetan Plateau. *J. Geol.* **113,** 517–533 (2005).
30. Graham, S. A. *et al.* Stable isotope records of Cenozoic climate and topography, Tibetan Plateau and Tarim Basin. *Am. J. Sci.* **305,** 101–118 (2005).
31. LePichon, X., Fournier, M. & Jolivet, L. Kinematics, topography, shortening and extrusion in the India–Eurasia collision. *Tectonics* **11,** 1085–1098 (1992).
32. Friedman, I. & O'Neil, J. R. in *Data of Geochemistry* (ed. Fleischer, M.) 1–12 (US Geological Survey Prof. Pap. 440-K, Washington DC, 1977).

Acknowledgements Particular thanks are extended to D. Schrag for use of his laboratory at Harvard University for the oxygen and carbon isotopic analyses. We thank M. Haizhou of the Salt Lake Institute, Chinese Academy of Sciences, Xining, for logistical assistance and arranging access to sites in northern and central Tibet. We also thank A. Cyr for assistance in the field. D.B.R. thanks fellow members of the Earth System Evolution Program of the Canadian Institute for Advanced Research for discussion and comments. This work was funded by the National Science Foundation.

Author Information Reprints and permissions information is available at npg.nature.com/reprintsandpermissions. The authors declare no competing financial interests. Correspondence and requests for materials should be addressed to D.B.R. (drowley@uchicago.edu).