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**Timeline of the South-Tibet--Himalayan belt: the geochronological
record of subduction, collision, and underthrusting from zircon and
monazite U-Pb ages**

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19 **Abstract**

20 The “exact timing” of collision between India and Eurasia is a recurring. Careful dating is critical
21 for all tectonic events. With the example of the South Tibet-Himalaya collision system, a short
22 review of arguments from different approaches suggests that this pursuit is in vain, but that our
23 knowledge is already sufficient to provide an acceptably “precise” timing of the main events. We
24 reviewed U-Pb ages of zircons and monazites considering that major tectonic events can produce
25 thermal effects strong enough to be recognized in high-temperature geochronology. This review
26 also shows that precise timing is beyond the precision of the methods and the rock record. General
27 consistency between geologic and thermochronologic records strengthens previous interpretations
28 of the collisional orogenic system. At variance with most tectonic interpretations, we argue that
29 the Tsangpo Suture in South-Tibet results from two merged subduction zones, and that island arcs
30 may be part of the root of the Eurasian paleo-active margin. The two main collisional events
31 closely followed each other at ca. 50 and 40 Ma.

32
33 Key words: Subduction, collision, India, Himalaya, U-Pb thermochronology.

35 **Introduction**

36 The discovery of plate tectonics in the late 1960’s allowed renewed interpretation and
37 classification of recent and ancient mountain belts (e.g., Dewey and Bird 1970; Dewey and Burke
38 1973; Dewey et al. 1973). Earth scientists learned that mountain ranges are built along convergent
39 plate margins by successive subduction, occasionally obduction and ultimately continental
40 collision (e.g., Cox and Hart 1986; Kearey et al. 2009; Park 1993), each process overprinted the
41 previous ones. The mega-annual tectonic sequences include magmatic, metamorphic and crustal

shortening-thickening processes (e.g., Condie 1997). Celal Şengör, one of the front-runners in the development of orogeny-plate tectonics relationships, demonstrated the three-dimensional consequences of pre-collisional irregularities on the strain distribution within mountain belts and their forelands (Şengör 1976). Many studies since then have shown that continental fragments, island arcs, and seamounts (as well as post-orogenic faulting and segmentation) introduce complexity in the collision zones (e.g., Şengör et al. 2018). Continental collision and subsequent suturing are therefore “capricious” and “diachronous” records of multiple tectonic events (Dewey 1977; Dewey and Kidd 1974). Although this assertion sounds like a late 1970’s truism, many authors have disputed the timing of collisional inception in various orogens, especially the Tibet-Himalaya plateau and mountains resulting from the first contact between India and Eurasia (e.g., Aitchison et al. 2007; Butler 1995; Rowley 1996). In the context of this example, we argue that the recurring controversy is pointless. Different methods yield different lines of evidence and age brackets that refer to different norms and constrain different processes. Combined, these methods yield bulk uncertainties that are, on a geological time scale, sufficiently narrow to satisfy our understanding of the initial collision, although the finest of details have likely forever escaped any geological record.

Studies of the South Tibet-Himalaya orogenic system has strongly influenced our general understanding of collisional orogens, though, for the same reason that different methods use distinct criteria, the suite of thermodynamic mechanisms involved in mountain building remains unclear. Here, we first attempt to synthesize the collisional history from geophysical and geological arguments. This review can only refer to a few of the hundreds of articles published on the topic, and confirms that continental subduction followed by intracontinental underthrusting of the Indian continent are the fundamental processes that produced the South-Tibet-Himalaya orogenic belt.

To determine whether the flurry of published geochronological measurements could deliver a more subtle time line to constrain the tectonic evolution of the orogenic system, we compiled high-temperature records from U-Pb dating of magmatic and metamorphic zircons and monazites. We avoided methods such as Rb-Sr, $^{40}\text{Ar}/^{39}\text{Ar}$ and fission tracks as are often linked to exhumation processes that may or may not be due to late tectonic events (e.g., Huntington et al. 2007). The results consistently show that southward-migrating crustal thickening and associated prograde metamorphism are Eocene to Early Miocene in age (ca. 40-20 Ma, e.g., Yin 2006, and references therein) and that out-of-sequence events such as magmatism in the North Himalaya domes and backthrusting along the suture zones remain to be understood.

Setting – Large-scale structure

The ~ 2500 km long, 250-350 km wide South Tibet-Himalaya belt is grossly described as an orogenic wedge formed in sequence in front of and including suture zones that bound the Eurasian backstop. As is apparent in early geological maps (e.g., Argand 1924; Gansser 1964) the belt has a relatively simple, arcuate and cylindrical geometry over most of its length and terminates at both ends in nearly transverse syntaxes named after the main peaks that tower above them: the Namche Barwa (7756 m) in the core of the eastern syntaxis and the Nanga Parbat (8138 m) in the western syntaxis in Pakistan (Fig. 1). Southwards-directed and -propagating thrusting has assembled South Tibet and the Himalayas from four main tectonic and lithologic zones that are remarkably consistent all along the length of the mountain belt (Fig. 1). From north to south, i.e., from the structural top to bottom (Fig. 2) these are:

- 1) The Eurasian paleo-active margin, so-called the Transhimalaya or Gangdese Batholith in South Tibet, which continues into the Karakoram Batholith in the western Himalayas.

2) The Tsangpo Suture Zone delineates the plate boundary along which the Tethyan oceanic lithosphere (ophiolites and deep-sea sediments) that separated India and Eurasia was subducted beneath Tibet. The suture includes tectonic slices of oceanic arcs in South Tibet, which are equivalent to the matured Ladakh and Kohistan Arcs in the western Himalayas (Fig. 1) where the Tsangpo Suture is split into the so-called Shyok Suture to the north and the Indus Suture to the south, between the arcs and India.

3) The Indian continent comprises three sub-units:

- the Tethys Himalaya exposes sediments deposited on the North Indian passive margin and its continental rise, which have been intruded by Eocene and younger granitoids in the North Himalaya domes;
- the High Himalaya mainly composed of ortho- and paragneisses intruded by Miocene leucogranites, is due to intra-continental underthrusting of peninsular India along the basal Main Central Thrust (MCT), and;
- the Lower Himalaya, an imbricate thrust pile of mid-Proterozoic to Cenozoic sequences pertaining to the Indian continental crust.

4) The Sub-Himalaya foredeep, south of the mountain belt.

Each zone is bounded by crustal fault zones (Fig. 2) and has a distinctive stratigraphy and metamorphic-magmatic history. They have been extensively and even repeatedly described with various names in many regional and review publications (e.g., Hodges 2000; Yin and Harrison 2000). Therefore, we will only refer to selected, pertinent information, and detailed descriptions can be found in the abundant literature. We will briefly summarize the tectonic framework of the South-Tibet-Himalaya collision system, discuss the timing of collision, and synthesize existing U-Pb geochronological data to explore how they further constrain important geodynamic events.

110 **Tectonic history**

111 Magnetic anomalies of the Atlantic and Indian Oceans and paleomagnetic measurements in Asia
 112 and India have revealed the relative movement of cratonic India relative to cratonic Asia since the
 113 Early Jurassic (e.g., DeMets et al. 2010; Klootwijk et al. 1992; Ricou 1994; Savostin et al. 1986;
 114 Torsvik and Cocks 2017).

115 *Tethys Ocean: rifting and ocean spreading*

116 A long period of left-lateral, tensional wrenching separated the northern (Laurasia) and southern
 117 part (Gondwana) of Pangea (e.g., Muttoni et al. 2003). Lithological, paleontological, geochemical
 118 and geochronological evidence from basement rocks and sedimentary cover indicates that India
 119 was part of Gondwana. Beginning in the late Paleozoic, rifting progressed westward along the
 120 northern Gondwana margin, opening space for the (Neo-)Tethys Ocean (e.g., Stampfli and Borel
 121 2002). During the Middle/Late Jurassic and Early Cretaceous, Gondwana split into a western
 122 (Africa and South America) and eastern landmass (Australia, Antarctica, India and Madagascar).
 123 These two landmasses fragmented further into their respective components during the Cretaceous.
 124 India separated from Antarctica and Australia in the Early Cretaceous (ca. 130-135 Ma) and from
 125 Madagascar in the Late-Cretaceous (85-90 Ma; Gibbons et al. 2013; Torsvik et al. 2000).

126 *Tethys Ocean convergence: subduction to collision*

127 Ocean-floor magnetic anomalies offer an external constraint on the convergence history. Although
 128 different models yield slightly different histories (e.g., van Hinsbergen et al. 2011), they provide a
 129 consistent, first-order image of the relative motion between India and Eurasia.

- 130 - Cratonic India began to drift northward toward cratonic Asia in the Late Cretaceous (100-
 131 90 Ma) associated with an anticlockwise rotation of India and representing about 8000 km
 132 of convergence (Patriat and Achache 1984; van Hinsbergen et al. 2012). Plate

reconstructions predict ~ 7500 and 8700 km of convergence at the longitudes of the western and eastern ends of the Himalaya, respectively (e.g. van Hinsbergen et al. 2011), indicating the extent of lithosphere consumed (oceanic Tethys and continental north India). On average, the displacement of India was rapid, reaching ~ 18 cm/year around 60 Ma (Patriat and Achache 1984; van Hinsbergen et al. 2012).

- Comparison of paleomagnetic data from South Tibet and apparent polar wander paths for the Indian plate shows that the collision between India and Asia occurred at equatorial latitudes, with progressive suturing since the Paleocene in the westernmost Himalaya (at 67-60 Ma; Beck et al. 1996; Smith et al. 1994) and until the Early Eocene (ca. 50 Ma) in the eastern Himalaya (Rowley 1996).
- India-Eurasia convergence began to decelerate exponentially around 50 Ma, usually attributed, along with short-lived changes in drift direction, to the onset of continental subduction and India-Eurasia collision (Patriat and Achache 1984). The displacement rate decreased to 4-5 cm/year by about 40 Ma and remained nearly steady until further deceleration around 20 Ma (Molnar and Stock 2009). The present-day rate of convergence across the Himalaya is 1.5-2 cm/year, just under half the total convergence rate (Ader et al. 2012; Bilham et al. 1997).

These 3-episodes of relative motion are generally geodynamically interpreted as follows:

- Subduction of Tethyan oceanic lithosphere. Pulled by an increasingly long subducting slab, the rate of displacement of India was enhanced (e.g., Schellart 2004). Push by the Réunion plume (Cande and Stegman 2011; van Hinsbergen et al. 2011) a positive feedback between two slabs (Jagoutz et al. 2015), or a long-lived convective cell (Sternai et al. 2016) may have contributed to the great velocities.

- Subduction of the Indian continental lithosphere. Deceleration of the relative convergence is attributed to the buoyancy of the continental lithosphere, which increasingly resisted subduction as the volume of the subducted continental crust increased (e.g., Molnar and Gray 1979).

- Intracontinental shortening. The steady, 4-5 cm/year average convergence rate is controlled by the deformability of the colliding continental lithospheres and potential pulling by the subducting Indian continental lithosphere (Capitanio et al. 2010). The strength and heterogeneities of the deforming continents control the deformation regime. The onset of strike-slip-dominated escape tectonics (Tapponnier et al. 1982) may explain the subsidiary 20 Ma slowing of convergence. Paleomagnetic data (Klootwijk et al. 1986) and differential convergence rates between the western and eastern edges of India (Molnar and Stock 2009; van Hinsbergen et al. 2011) indicate ongoing anticlockwise rotational underthrusting of India in the Himalaya.

Subduction to collision

The Kohistan and Ladakh Arc terranes in the NW Himalaya indicate that intra-oceanic subduction was well established by ca. 135 Ma (e.g., Burg 2011; Rolland et al. 2000; Thakur and Misra 1984), and may even have started earlier, around 155 Ma (Jagoutz et al. 2018). These arcs, active throughout the Cretaceous, are remnants of a chain of island arcs that extended eastward, possibly for the entire length of the India-Asia suture zone (Burg 2007; Hébert et al. 2012). Such remnants are also found within the strike-slip-dominated lateral boundaries of the Indian Plate e.g., the Muslim Bagh ophiolites along the Chaman Fault (Kakar et al. 1971) and the Naga Hills ophiolites in the Indo-Burman range (Dey et al. 2018). To the north of these volcanic arcs, the back-arc oceanic lithosphere was subducting below the Eurasian active margin (i.e., the Transhimalaya and Karakoram terranes; Fig. 1), producing the abundantly dated calc-alkaline magmatism (e.g. Searle et al. 2010; Wen et al. 2008). A two-subduction system was therefore consuming the large Tethys

179 Ocean at least until ca. 95 Ma (Burg 2007, 2011), and may have ceased upon the 90-85 Ma plate
180 reorganization and the separation of India from Madagascar. “Himalayan” blueschists exhumed at
181 that time (Anczkiewicz et al. 2000; Honegger et al. 1989) may also represent this global event.
182 Granodioritic and volcanoclastic pebbles in Aptian-Albian conglomerates of the forearc of the
183 Transhimalaya indicate that the Asian active margin was already deeply eroded at that time,
184 suggesting an elevated Andean-type orogen since the Cretaceous (An et al. 2014; Burg and Chen
185 1984; Einsele et al. 1994). Besides magmatic and geochemical evidence, we know very little about
186 Early Cretaceous subduction. Recent tectonic reconstructions suggest that the oceanic arc was then
187 at equatorial latitudes (Hafkenscheid et al. 2006). Positive P wave seismic velocity anomalies in
188 the mantle beneath South India may represent sunken slabs of this intra-oceanic subduction (Van
189 der Voo et al. 1999; Zahirovic et al. 2012).

190 Magmatic products and their geochemistry are further evidence for Late Cretaceous to Eocene
191 subduction (e.g., Bouilhol et al. 2013; Ji et al. 2009), and plate reconstructions agree on the
192 northward drift of India and coeval consumption of several thousand kilometres of oceanic
193 lithosphere (e.g., Besse and Courtillot 1991; Schettino and Scotese 2005; Stampfli and Borel 2002;
194 Torsvik and Cocks 2017). The 80 to 70 Ma hiatus in dated subduction-related magmatism of the
195 Transhimalaya, Andean-type margin may reflect a change in slab behaviour, be it cessation of the
196 subduction that closed the back -arc basin or roll-back or breakoff (Ji et al. 2009; Wen et al. 2008).

197 **Age of collision initiation**

198 *What defines collision initiation?*

199 The initiation of the India-Eurasia collision should refer to the first contact between Indian and
200 Eurasian continental crusts after disappearance of the Tethyan oceanic lithosphere. The age of this

event is still disputed between disciplines. In this discussion, it is worth recalling key results used to set the framework in which collision took place.

Paleomagnetic and oceanic magnetic anomalies

Perhaps one of the most influential attempts to date the collision between India and Eurasia has been the study of India's motion with respect to the more stable Eurasian plate using seafloor magnetic anomalies (Molnar and Tapponnier 1975). They revealed a sharp decrease in convergence rate in the Eocene, which was interpreted as "the most likely time for initial contact between India and Eurasia". Higher resolution data and paleomagnetic interpretations confirmed this rapid deceleration and revealed accompanying erratic shifts of India's motion between anomalies 22 and 18 (50-42 Ma; Patriat and Achache 1984). These rate and directional changes were attributed to a fluctuating balance between pulling and buoyancy forces during subduction of the irregular northern continental margin of India. Later reconstructions of India relative to Eurasia (e.g., Besse et al. 1984; Copley et al. 2010; Dewey et al. 1989; Klootwijk et al. 1992; van Hinsbergen et al. 2011) agree that oceanic subduction continued until ca. 50 Ma (anomaly 22) and was followed by increased resistance to subduction until ca. 44 Ma (anomaly 20). This period of "soft" collision changed to "hard" collision when buoyant forces of the underthrust continental crust opposed further subduction, beginning the ongoing continental shortening and coeval thickening in both India and Eurasia (Chemenda et al. 2000). Whether separate accretionary events and crustal deformation episodes, including the Himalayan orogeny, may be correlated with (or responsible for) episodic pulses in sea floor spreading at the trailing edges of India (e.g., Merkouriev and DeMets 2006; Molnar and Stock 2009; White and Lister 2012) remains debatable because the uncertainties associated with each reconstruction are not resolved.

223 Stratigraphic and sedimentary criteria

224 In general, oceanic closure due to continental collision is correlated with a change from marine to
225 continental sedimentation and terrestrial fauna trespassing the ophiolitic suture.

226 The oldest reported faunal exchange between India and Eurasia is that of amphibians between
227 western India and Afghanistan (Jaeger et al. 1989), suggesting that continental bridges between
228 the two continental blocks at the western end of the Himalayas existed as early as the Cretaceous-
229 Paleocene boundary (around 65 Ma). The ca. 54 Ma exchange of terrestrial fauna between India
230 and Asia is similarly consistent with early continental pathways between the two continental
231 masses (Clementz et al. 2011; Kapur et al. 2017). These faunal migrations have been linked to the
232 initiation of the collision. However, the western margin of the Indian plate was a transform
233 boundary along which promontories may have formed early connections without physical
234 consequences on the India-Asia collision in South Tibet. The first arrival on the Indian plate of
235 clastic grains attributed to the active margin of Eurasia is dated at about 52 Ma (Najman et al.
236 2010; Zhuang et al. 2015). Yet, detritus reaching India may have been far-travelled if the trench
237 had been filled at that time, and a western provenance cannot be excluded. Furthermore, Early
238 Cretaceous (ca. 140-115 Ma) extension-related dykes, sills, and magmatic stocks intruded Tethyan
239 sediments (Chen et al. 2018; Zhu et al. 2009), and Cretaceous clastic zircon grains found in Eocene
240 sediments may indicate that such intrusions were already eroded at the time of deposition.

241 Paleostratigraphic evidence dates the development of a flexural bulge in the western (Ladakh)
242 outer margin of India to the Paleocene-Eocene boundary, ca. 55 Ma (Garzanti et al. 1987). This
243 bulge implies that northwestern continental India had entered the trench and begun subducting at
244 that time. Marine, shelf sedimentation on the northern continental margin of India continued until
245 the Eocene in South Tibet (Mu et al. 1973), and more precisely into the Priabonian (38-34 Ma,

Jian et al. 2016). This age is almost 10 m.y. younger than the final marine deposits preserved in other Himalayan regions (short review in Jian et al. 2016) with a possible tendency of slightly older ages in the western regions that would support the west to east collisional progression during the Eocene (Rowley 1996) due to anticlockwise rotation of India (Patriat and Achache 1984). In South Tibet, an abrupt change in lithology and clastic content around 60 Ma is attributed to the earliest contact between Asia and continental India, which was followed by southward migration of the Indian continental bulge (DeCelles et al. 2014).

The so-called Xigaze Basin, an important element of the suture, is a forearc basin with Early Cretaceous deep marine radiolarites deposited atop ophiolites (Allègre et al. 1984); overlying turbidites followed by Late Cretaceous shelf and fluvio-deltaic marine strata witness the filling of the basin with no sign of early collision up to ca. 65 Ma. (An et al. 2014; Einsele et al. 1994; Hu et al. 2016). Adjacent to the forearc sequences, non-marine Upper Oligocene-Lower Miocene (Kailas) conglomerate-rich sequences unconformably cover the Eurasian magmatic arc. They are ascribed to an extensional basin conjecturally related to southward rollback of the Indian continental lithosphere, ca. 30 m.y. after the initiation of collision (DeCelles et al. 2011). Reconciling the various sedimentary settings, the transition from marine to continental facies must have occurred during the Eocene.

Tectonic and deformation criteria

Discussing the time of collisional initiation assumes uninterrupted Indian continental crust entering in contact with the continuous Eurasian lithosphere. However, such an assumption is hardly tenable when considering the Tethyan paleogeography. Several realms come in consideration.

The northern continental margin of India may well have been an extended margin with several continental ribbons and fragments separated from the main continent. Such “allochthons” are

common in distal parts of modern margins (e.g., Brune et al. 2014; Mohn et al. 2015). They may have been the basement of the distinctive Permian reef limestones found as exotic blocks in the South Tibet olistostrome (Jin et al. 2015) and other Tethyan regions from Turkey to the Himalaya (Marcoux and Baud 1996; Stampfli et al. 1991). This possibility questions the nature, extent and thickness of the missing lithosphere discussed as “Greater India” (e.g., Ali and Aitchinson 2005; van Hinsbergen et al. 2012)

The Tethys Ocean may have contained seamounts (Dai et al. 2012) and a chain of island arcs forming the eastern correlatives of the Kohistan-Ladakh arc (Burg 2007; Hébert et al. 2012). Related back-arc and forearc basins had unknown widths, possibly variable along strike. These lithospheric irregularities have been accreted to the southern margin of Eurasia. Each accretionary event would qualify as collision initiation and may be apparent as the reported erratic behaviour of India (e.g., White and Lister 2012).

281 Metamorphic criteria

Eclogite facies and ultra-high-pressure (UHP) metamorphic assemblages of mafic rocks belonging to continental India have been dated to constrain the collisional timing and processes. These few occurrences in the western Himalaya indicate that the leading edge of the Indian crust was subducted down to ~100 km (metamorphic pressure > 3 GPa) at 55-45 Ma (de Sigoyer et al. 2000; Guillot et al. 2008; Kaneko et al. 2003; Leech et al. 2005). A simple trigonometric calculation shows that with an average Indian plate velocity of 4.5 cm/year (Besse et al. 1984; Klotwijk et al. 1992; Molnar and Stock 2009) and a ~ 20° subduction angle consistent with the mantle wind (Ficini et al. 2017), the given age would date the arrival of continental India in the trench at about 70 Ma. The subduction angle may have been steeper (ca. 40°) later (Guillot et al. 2007); any estimate assumes large uncertainties. Rapid exhumation of these rocks followed a retrograde path

292 through amphibolite conditions between 45 and 35 Ma to near-surface cooling at ca. 25-5 Ma (e.g.,
 293 Wilke et al. 2001).
 294 Granulitized eclogites in the High Himalaya (Burg et al. 1987; Grujic et al. 2011; Lombardo et al.
 295 2016) record pressures > 1.5 GPa (50-60 km depth) before near isothermal ($\sim 700^\circ\text{C}$)
 296 decompression to < 0.4 GPa. These eclogites have been variously dated to 40-23 Ma with re-
 297 equilibrated in the granulite facies during the Middle Miocene (15-10 Ma; Corrie et al. 2010; Cottle
 298 et al. 2009; Kellett et al. 2014).

299 Geochemical and isotopic criteria

300 The Transhimalayan magmatic belt formed the active continental margin of Eurasia during the
 301 north-dipping Tethys subduction (e.g., Allègre et al. 1984). Whole rock Sr and Nd isotope ratios
 302 and zircon Hf isotopic compositions indicate a dominantly juvenile mantle source of the mostly I-
 303 type granitoids (Harris et al. 1988). Negative ϵHf and ϵNd values in related volcanic rocks indicate
 304 the Gondwana affinity of the pre-batholith, Archean-Proterozoic continental basement (Zhu et al.
 305 2011). Compiling zircon U/Pb ages (Fig. 3) shows that the continental margin of Asia (Gangdese
 306 and Karakoram Batholiths, Fig. 1) was magmatically active from the Jurassic (ca. 190 Ma, Chu et
 307 al. 2006) to the Miocene (up to ca 9 Ma, Zhang et al. 2014). Jurassic magmatism was possibly
 308 related to south-dipping subduction of the Bangong-Nujiang oceanic lithosphere to the north of
 309 this Gondwana-derived continental block (e.g., Zhu et al. 2011), but alternatively may be related
 310 to early north-dipping subduction of the Tethys Ocean (e.g., Kang et al. 2014). Cretaceous- Eocene
 311 calc-alkaline plutons (forming the batholith) are admittedly attributed to this Tethys subduction
 312 (Allègre et al. 1984). Peaks in the zircon age distribution (Fig. 3) should be cautiously interpreted
 313 in terms of large magmatic volumes or flare-ups because they often denote the number of
 314 geochronological data measured on similar samples or plutons, as near Lhasa in the Gangdese

(e.g., Wen et al. 2008) and near the Karakoram Fault (e.g., Ravikant et al. 2009). Nonetheless, the existing data suggest quiescence between Cretaceous (120-80 Ma) and Paleogene (70-45 Ma) stages (Wen et al. 2008). Even accounting for the time of magma ascent, the Eocene ages have been taken as evidence for cessation of subduction and disappearance of oceanic lithosphere between India and Asia ca. 45 Ma (Allègre et al. 1984). Changes in ϵ_{Hf} values of magmatic zircons are ascribed to slab breakoff ca. 52 Ma (Zhu et al. 2011). However, owing to bias in the data set, and because the southern Eurasian margin developed on a Gondwanian basement, changes in isotopic compositions of the Transhimalaya Batholith cannot be definitely ascribed to changing subduction dynamics and collision (Bouilhol et al. 2013). Younger (30-9 Ma), protracted magmatic episodes are ascribed to lithospheric phenomena after closure of the Tethys Ocean (e.g., Zhang et al. 2014).

In contrast to the Transhimalaya Batholith, the Kohistan–Ladakh plutonic rocks developed from juvenile magmatic activity in oceanic lithosphere since at least ca. 135 Ma and until ca. 40 Ma (Fig. 3). In these oceanic arcs, compositional changes may be interpreted in terms of subduction dynamics. In-situ U–Pb geochronology, zircon Hf isotopic compositions, and whole-rock Nd and Sr isotopic data indicate that the Kohistan–Ladakh island arcs were accreted to India around 50 Ma and that the India/arc assemblage collided with Eurasia around 40 Ma (Bouilhol et al. 2013; Garzanti et al. 1987).

Collision – orogenic wedge

Sutures

Multiple collisions are evident where the Kohistan and Ladakh arcs are preserved between two sutures (as adopted in Fig. 4). In South Tibet, only the Tsangpo Suture (Fig. 1) is delineated by a set of dismembered ophiolites representing the Tethys Ocean, fore-, inter-, and back arcs (Hébert

et al. 2012), and small remnants of oceanic arcs (Aitchison et al. 2000). This diversity implies that the Tsangpo Suture zone encompasses and overprints several subduction and collision events (Burg 2007). After some dispute as to whether the island arcs were accreted to Eurasia during the Cretaceous (e.g., Coward et al. 1982; Frank et al. 1977; Petterson and Windley 1985) or only after accretion to India in the Eocene (e.g., Bard 1983; Klootwijk et al. 1979), geochronological and isotopic studies on plutonic rocks of the Kohistan-Ladakh arcs linked changes in the magmatic source regions to arc/India collision at ca. 50 Ma followed by Eurasia/arc collision at ca. 40 Ma (Bouilhol et al. 2013). Such data do not preclude a more complex history in which several thousand kilometres of Tethys Ocean and its various (perhaps unrelated along-strike) oceanic plateaus, forearcs, arcs and back arcs were reduced to a narrow suture a few metres to few kilometres wide and several thousand kilometres long in geological maps. Overprinting by late faulting such as Miocene backthrusting bounding the southern Transhimalaya (Burg and Chen 1984; Yin and Harrison 2000) and the Karakoram Batholith (Searle et al. 1988) may have buried the accreted units, rendering the successive discrete collisions impossible to decipher. Such buried units may provide an explanation for “bodies” of seismic reflectors observed along the suture zone (Guo et al. 2017; Makovsky et al. 1999). Overall, the timeline suggests a general northward migration of collisional events until hard, total closure ca. 40 Ma, when the never-subducted Xigaze forearc of the Transhimalaya was folded (Burg and Chen 1984; Yin and Harrison 2000).

Greater India

Greater India, i.e., the pre-collisional extent of India’s northern continental crust, is difficult to outline because of large uncertainties in the locations and shapes of pre-collisional margins destroyed by subduction and collisional deformation. However, the extent of Greater India is a critical question since it directly addresses how and where collision began to affect the continent

and how subsequent convergence between India and Asia may have been accommodated. The proposed extent of Greater India varies by up to several thousand kilometres according to different authors (review of Ali and Aitchinson 2005). Several tectonic reconstructions fitting the possible northern boundary of Greater India with the margins of Australia suggest around 800 km of continental crust to the north of the Himalayan front (e.g., Zahirovic et al. 2012) possibly preceded by up to ~2000 km of Indian oceanic lithosphere (e.g., van Hinsbergen et al. 2012). Such a wide passive margin may have been strongly extended (thinned), and would have been easily subducted without a noticeable effect on the plate motion. With such freedom, we have no reason to depart from most of the aforementioned interpretations. We can accept a set of pre-50 Ma events related to various small collisions with oceanic plateaus, arcs, and continental allochthons of the thinned passive margin (Greater India) while slab pull remained dominant and contributed to the very high convergence rate. At about 50 Ma, the main Indian continental crust entered subduction along with island arcs, an event sufficient to drastically slow Indian plate motion. Since buoyancy forces were subordinate until that time, the subduction system (soft collision) was essentially ablative (non-accreting sensu von Huene and Scholl 1991). Hard collision, when buoyancy forces became an important and caused accreting orogeny, began at that time.

Slab break-off

Detachment of the oceanic slab from the continental Indian lithosphere has been considered as an efficient mechanism to interrupt pull (van Hunen and Allen 2011), hence slowing the incoming Indian plate between 50 and 40 Ma and changing the collisional dynamics from soft to hard collision (Chemenda et al. 2000). Several slab break-off events during the India-Eurasia convergence have been inferred on the basis of secular changes in magmatic arc composition, in particular the ca. 50 Ma magmatic flare-up of shoshonitic magma in the Transhimalaya (Lee et al.

2009) and termination of arc-related magmatism shortly thereafter, around 45 Ma (Chung et al. 2005). As previously discussed, relating compositional changes in Transhimalaya plutonic rocks to subduction dynamics is uncertain since crustal thickness may affect melt composition, especially in continental magmatic arcs (Freeburn et al. 2017). Slab break-off interpretations were motivated by at least two low-velocity anomalies in the lower mantle, which may be attributed to stagnant slabs of Tethyan oceanic lithosphere beneath India (Replumaz et al. 2010; Van der Voo et al. 1999). Analogue (Chemenda et al. 2000) and numerical (Magni et al. 2017) models relevant to the India-Asia collision suggest that the inferred Eocene loss of oceanic lithosphere by slab break-off during continental subduction could trigger exhumation of UHP metamorphic rocks of the Indian plate, which in turn underplating the overriding Asian plate.

Accreting collision

The three sub-units of the exposed northern margin of India (from north to south the Tethys, High and Lower Himalaya; Figs. 1 and 2) represent the south-directed collisional accretionary system that developed in sequence at the expenses of “Greater” India. Some authors have suggested that collision may have started earlier in the west and subsequently propagated eastward along the Himalaya (e.g., Klootwijk et al. 1986; Rowley 1996). Such a diachronous, eastward migrating collision is consistent with faster convergence rates in the eastern corner of India than of the western corner (e.g., Molnar and Stock 2009; van Hinsbergen et al. 2012).

In the western Himalayas, the direct footwall of the suture includes gneisses and UHP eclogites dated at 55-50 Ma (de Sigoyer et al. 2000; Kaneko et al. 2003). In South Tibet, 40 Ma high-pressure gneisses are in a similar structural position (Laskowski et al. 2016). These rocks demonstrate that continental material was buried to about 100 km in the Eocene. As already discussed, these high-pressure metamorphic ages reflect the arrival of continental, Indian material in the trench around

407 70 Ma, assuming an average Indian plate velocity of 4.5 cm/year and a $\sim 20^\circ$ subduction angle are
 408 assumed. Yet, this information does not stipulate whether India was subducting beneath the
 409 Kohistan-Ladakh and laterally equivalent island arcs or beneath the Karakoram-Transhimalaya
 410 margins of Eurasia.

411 In South Tibet, isoclinally folded turbidites unconformably covered by the latest-Cretaceous-
 412 Paleocene olistostrome represent the oldest thrust system adjacent to the suture; they are
 413 interpreted to be part of an accretionary wedge (Burg and Chen 1984). Provenance analysis of
 414 clastic zircons in sandstone blocks suggests both Cretaceous and Late Paleocene-Early Eocene
 415 deposition (An et al. 2017). Cretaceous deposition occurred during an essentially non-accreting
 416 subduction, which we interpret to be the trench in front of the island arcs, later amalgamated with
 417 the Paleocene/Eocene wedge in front of the Transhimalaya forearc (Fig. 4). These rock units were
 418 later reformed in the Late Eocene, along with the north- and south subdomains of the Tethys
 419 Himalaya (Burg and Chen 1984). Since the Xigaze forearc was folded in the same time interval,
 420 hard collision was well in progress during the Late Eocene.

421 Further south, the High Himalayan gneiss consists of a north-dipping pile of tectono-stratigraphic
 422 units that, according to isotopic data, are late Proterozoic to early Paleozoic rocks of the upper
 423 continental crust of India (e.g., Parrish and Hodges 1996; Whittington et al. 1999). Mesozoic
 424 sediments, originally situated on the northern margin of India, lay locally on this monotonous
 425 gneiss sheet that varies in thickness from 2 to 10 km (e.g., Le Fort 1975). This thrust sheet is a
 426 deeply rooted crustal slice (Nelson et al. 1996) whose bottom is the ductile MCT. The gneiss and
 427 migmatites bear evidence of multiple deformation and recrystallisation events ascribed to
 428 Himalayan southward shearing, with the dominant fabric defined by amphibolite facies
 429 assemblages (e.g., Brunel 1986). The High Himalayan gneiss was intruded by Miocene

leucogranites contemporaneous with normal faulting (Burg et al. 1984a). The north-dipping extensional system known as South Tibetan Detachment (Burchfiel et al. 1992) is the tectonic contact separating the High Himalaya from the Tethyan sedimentary sequences along the mountain belt. The following analysis of thermochronological data focuses on whether those data allow refining of the tectonic narrative.

Thermochronological record in Indian rock units

To better constrain the high-temperature timing of the different orogenic zones, we have compiled U-Pb data of zircons from magmatic rocks in domes of the Tethys Himalaya (Fig. 5), and U-Pb zircon ages and U-Th/Pb monazite ages of the High Himalayan gneiss (Fig. 6). Recent advances in linking monazite U-Th/Pb ages with their microstructural setting and rare earth element concentrations allow distinction between prograde and retrograde grains (e.g., Pyle and Spear 2003). We have selected published U-Th/Pb data from transects across the High Himalayan gneiss, carefully identifying “prograde” high-Y from “retrograde” low-Y monazite ages. In general these thermochronological data are consistent with the overall southward propagation of magmatic and metamorphic events.

Tethys Himalaya

In South Tibet, a string of domes (collectively called the North Himalayan Belt) stretches about 50 km south of, and nearly parallel to the Tsangpo Suture, within the Tethyan Himalaya (Fig. 1). They are culminations of a long, orogen-parallel antiform attributed to ramp-thrusting and subsequent extension, diapirism, or a combination of these mechanisms (Lee et al. 2000). These domes expose Proterozoic (up to ca. 1800 Ma inherited zircon cores, e.g., Aoya et al. 2005; Liao et al. 2008) and Early Paleozoic (Lee et al. 2000; Schärer et al. 1986) basement orthogneiss intruded by small-volume granitoids (e.g., Aikman et al. 2008). Our zircon age compilation shows

rather continuous melt production with five main pulses occurring almost rhythmically every ~10 m.y., from 45 to 7 Ma (Fig. 5, references in supplementary material). The gneissic-granitic domes core concentric greenschist to amphibolite facies metamorphic zones (Burg et al. 1984b; Lee et al. 2004) on which ^{40}Ar - ^{39}Ar dating yields cooling ages of micas at ca. 10 Ma and fission track apatite ages of ca. 5 Ma (Lee et al. 2000).

These metamorphic domes provide evidence of burial down to 0.7-0.9 GPa for 600-700°C at 50-45 Ma (e.g., Ding et al. 2016; Lee et al. 2004). Zircon populations separated from the orthogneiss and the granitoids show a wide range of inheritance, with a dominant proportion of Ordovician grains among younger grains spanning the Mesozoic (Fig. 5, references in supplementary material). The cores of those inherited Ordovician grains indicate that Indian orthogneisses were important melt sources, whereas the Mesozoic grains indicate that Tethyan lithologies contributed to the melts from which Eocene and Miocene granitoids were generated. The bulk-rock isotopic compositions of these granitoids show negative initial Nd isotopic ratios ($\epsilon\text{Nd}_i \sim -14$) and high radiogenic Sr ($^{87}\text{Sr}/^{86}\text{Sr} > 0.715$), consistent with such crustal source components (e.g., Gao and Zeng 2014; Zeng et al. 2011). However, the ca. 45 Ma intrusions in the North Himalayan domes and surrounding Tethyan sediments are mantle-derived ($\epsilon\text{Nd}_i \sim -6$, Ji et al. 2016a).

High Himalaya

The High Himalaya gneiss pile (Fig. 6) shows a multi-stage metamorphic history with metamorphic grade increasing structurally upward (e.g., Kohn 2014). The most prominent stages are:

- 1) Remnants of eclogite-facies metamorphism (≥ 1.5 GPa) within the gneiss show a wide age range between 40 Ma (Kellett et al. 2014; Laskowski et al. 2016) and as young as 15 Ma for those closest to the MCT (Corrie et al. 2010), which is at odds with the older ages of the

surrounding gneisses and migmatites. These eclogites were recrystallized during Miocene granulite facies metamorphism at $> 750^{\circ}\text{C}$ and ~ 1 GPa (Groppo et al. 2007; Kellett et al. 2014).

- 2) Barrovian-type metamorphism: The north-dipping gneiss pile above the MCT shows a northward, upward-decreasing metamorphic sequence from $\sim 550^{\circ}\text{C}$ near the MCT to 700°C about 5 km above in the overlying migmatites, with pressures consistent with a lithostatic gradient (e.g., Hodges et al. 1988; Pêcher 1989). This metamorphic event started its prograde path at 45-35 Ma, likely during burial of the Precambrian protolith following continental collision. Metamorphism may have started earlier (50-45 Ma, Smith et al. 1994) in the western than in the eastern Himalaya.

Conversely, the metamorphic sequence is inverted in the MCT zone, with a continuous downward decrease in metamorphic grade and temperature. The sillimanite-bearing migmatites and gneiss occur above a narrow zone of kyanite-bearing schists, in turn above successively staurolite, garnet, biotite, and lower grade metamorphic zones (Fig. 6). Within this MCT zone, synmetamorphic shearing evolved at 20-25 Ma. Zircon and monazite ages are younger (20-10 Ma, e.g., Mottram et al. 2015) than in the upper pile. Burial of the footwall rocks ceased at ca. 10 Ma but further reactivation lasted into Late-Miocene (5-8 Ma) times (5-8 Ma, e.g., Oliver et al. 1995).

- 3) The dominant metamorphic event occurred under higher temperature and lower pressure conditions ($650\text{-}800^{\circ}\text{C}$ and $0.4\text{-}0.7$ GPa) during the Miocene (30-18 Ma, Clarke et al. 2016), obliterating most of evidence of the earlier, higher pressure event.
- 4) Low-pressure and -temperature conditions are mostly recorded in the contact zones of the 25-12 Ma tourmaline-muscovite-garnet leucogranites that define a belt of small plutons,

stoks and dyke networks (e.g., Cottle et al. 2015). Their shear fabrics give evidence for normal faulting and rapid denudation in the upper levels of the High Himalaya (Burchfiel et al. 1992; Burg et al. 1984a). Residual and/or peritectic andalusite and sillimanite, and biotite inclusions in cordierite indicate that melts formed by dehydration melting of muscovite at 660-700 °C and ~0.7 GPa (Patiño Douce and Harris 1998). Further exhumation later in the Miocene (20-15 Ma) occurred at lower, sillimanite grade pressure and temperature (500-700°C and 0.2-0.4 GPa).

Monazite ages

Monazite rare earth element concentrations allow distinction between prograde and retrograde growth. More than half of the 668 selected U-Th/Pb ages from monazites of the High Himalayan gneiss pile are prograde (Fig. 7). Their age distribution is asymmetric: prograde monazites increase in occurrence from 40 to about 20 Ma before sharply decreasing from 20 to 16 Ma (Fig. 7). A precursory small peak around 36 Ma represents the upper part of the gneiss pile, whereas prograde monazite ages are younger structurally downward towards the MCT (Fig. 6, Kohn 2014). Peaks exceeding the distribution envelope at 17 and 20 Ma correspond to over-represented samples that we were not able to objectively filter from the dataset.

Retrograde monazite ages also show an asymmetric distribution: retrograde monazites increase in occurrence from 30 to 16 Ma and quickly decrease by 13 Ma (Fig. 7), the time of cooling below ~700°C (Parrish 1990). Like the prograde ages, the retrograde ages are younger structurally downward towards the MCT (Fig. 6).

Zircon ages

Only a few U-Pb zircon ages between 50 and 40 Ma have been reported from the High Himalayan gneiss pile (Fig. 7). Apart from those, the histogram envelope of zircon ages increases from 38 to

25 Ma in a similar asymmetric distribution as the prograde monazites, peaks between 25 and 20 Ma, then sharply decreases between 20 and 15 Ma. The abundant 25-15 Ma age population includes the time of crystallization of most High Himalaya leucogranites (Fig. 7, references in supplementary material). The peak departing from this general trend at 32 Ma possibly represents zircon dates from outcrops that were the subjects of several publications, though insufficiently reported sampling locations prevented us from filtering those data.

Discussion

Continental subduction of Indian continental margin

The Indian continental basement is exposed in three settings: (1) UHP gneisses, (2) the North Himalayan Domes and (3) the High Himalaya. In addition, many Indian-derived inherited zircon are found in the Kohistan-Ladakh and Transhimalayan plutonic rocks (Bouilhol et al. 2013; Chu et al. 2011). These zircons, along with the Hf isotopic constraints, indicate that the Indian plate was underthrust beneath these arc regions during the Eocene (Xu et al. 2010). The metamorphic age of the northernmost exposed UHP part of India is up to 55 Ma, and metamorphism and melting occurred at 50-40 Ma in the North Himalayan Domes and 40-20 Ma in the High Himalaya (Fig. 7). These ages are therefore consistent with the structurally inferred history of in-sequence, southward propagation of the South Tibet-Himalaya orogenic system.

North Himalayan Domes

Alkaline gabbros of mantle origin are present in the North Himalayan Domes and are contemporaneous with ca. 45 Ma crustal granitoids. These mantle melts are interpreted to represent slab break-off (Ji et al. 2016b), though 45 Ma corresponds to the age ascribed to the UHP metamorphism in the NW Himalaya and to retrogressed eclogites in the High Himalaya. Coeval mantle and crustal melting events are difficult to interpret in a continuous and uniform Indian

lithosphere, as is postulated to underlie the Tethyan sedimentary sequences. Discontinuities may be inherited from the structure of the passive margin of India. It is worth noting that the North Himalaya basement rocks crop out below Late Paleozoic series (e.g., Burg et al. 1984b; Lee et al. 2004) whereas Cambrian-Ordovician and younger Paleozoic sequences remain on the northern flank of the High Himalaya pile (e.g., Burg and Chen 1984; Liu and Einsele 1994). If the stratigraphy is correct, the South Tibetan Detachment cannot be a unique fault zone from the High Himalaya to the domes (Lee et al. 2000) because it would ramp up-section in the direction of movement. Instead, different stratigraphic layers may have been deposited at different times on low and high extensional blocks of the North Indian passive margin. Such structural discontinuities would have been prone to compressional reactivation during the Cenozoic.

Along with the metamorphic record in the domes, mantellic and crustal magmas suggest burial of the Indian crust during Eocene subduction and possible decompression melting of the underlying mantle when exhumation began. This interpretation questions the importance of the thrusts mapped to the south of the domes (e.g., Burg and Chen 1984), whose throw may have been underestimated. In any case, the geographical and age distributions in these domes do not support a double-sided, centripetal slab detachment (Webb et al. 2017) or southward migrating delamination of the Indian lithosphere (DeCelles et al. 2011). Should we include the UHP occurrences in the same group of basement outcrops as the North Himalayan Domes, they may all represent segments of blocks of the North Indian margin (Fig. 4) that underwent parallel but separate subduction histories before upward extrusion within the Tethyan realm.

High Himalaya

Early metamorphic/magmatic events in the High Himalaya Indian crust are indicated by the ca. 35Ma zircons in High Himalaya leucogranites (Fig. 5, e.g., Zhang et al. 2012), about 10 m.y. later

568 than early events in the North Himalayan Domes (Aikman et al. 2008). Consistent monazite and
 569 zircon ages suggest that thrusting deformation began in the Late Eocene.

570 The age distributions of zircons and prograde and retrograde monazites bracket the time of High
 571 Himalayan metamorphism. Prograde and retrograde monazite ages (Fig. 7) coincide with the ages
 572 of most High Himalaya leucogranites (e.g., Montomoli et al. 2015), suggesting that the
 573 metamorphic climax and main melting event of the High Himalaya gneiss pile occurred around 20
 574 Ma, as concluded in a number of previous studies (e.g., Harris et al. 2004; Singh 2018; Weinberg
 575 2016). The system then cooled rapidly within ca. 5 m.y., with downward strain localization and
 576 southward propagation of the increasingly brittle deformation accompanying thrusting of the High
 577 Himalaya on its foreland. The fact that older ages are statistically found in the structurally upper
 578 part of the gneiss pile and that prograde and retrograde ages are young downward is symptomatic
 579 of in-sequence Himalayan thrusting (e.g., Kohn 2014). Prograde ages in the footwall of the basal
 580 MCT and retrograde ages in the hanging wall (Kohn 2014) are consistent with post-12 Ma
 581 transport of the High Himalaya metamorphic sequence over the underlying Lower Himalaya units.

582 ***High-Himalaya-North Himalayan Domes conundrum***

583 These thermochronological arguments support the idea that the orogenic wedge developed grossly
 584 in sequence. The variety of magmatic and metamorphic ages reveals a varied tectonometamorphic
 585 history along the belt.

586 A part of the conundrum is that there is no plutonic crystallization age between 10 and 25 Ma in
 587 the North Himalayan Domes, though this spans the time of leucogranite intrusions on the top of
 588 the High Himalaya. Yet, there is “out-of-sequence” leucogranitic magmatism as young as ca. 9
 589 Ma in the domes (Fig 5). If there are 45 Ma crustal and mantle-derived intrusions in the domes,
 590 there is no other evidence for melting at this time in the High Himalaya other than melt inclusions

in kyanite-bearing gneiss dated at about 42Ma (Carosi et al. 2015) and few zircon ages. These provide the sole witness of a possible temporal link between an early phase of metamorphism and melting in the High Himalaya and the domes. Such thermochronological data advocate for thermal decoupling between the two orogenic settings, which has already been suggested from Ar/Ar dating (Coleman and Hodges 1998). These data also indicate that early melting of the Indian crust under high-pressure conditions may have contributed to the Eocene intrusions found in the North Himalayan Dome.

A working hypothesis would be, again, to consider the North Himalyan domes as outliers of rifted Indian margin (Fig. 4) that were accreted within the Tethyan sediments during the early stages (55-40 Ma) of continental subduction. The High Himalaya formed by in-sequence thrusting after about 30 Ma. Underthrusting of India allowed stacking and underplating of continental material via duplexing and extrusion to form the accretionary wedge as we know it today.

Conclusions

The record from different disciplines is sufficient to confirm the general framework of collisional tectonics in the South Tibet-Himalaya orogenic system.

The collision of India and Eurasia is marked by two main suture zones bounding island arcs in the western Himalaya. Tectonic shreds of island arcs along the Tsangpo Suture in South Tibet suggest similar protracted collisional events. Sometime in the early Paleocene, northern India, and possibly continental slivers ahead of it, collided with the chain of intra-oceanic island arcs at near-equatorial latitudes. The missing parts of the arcs and their back-arc lithosphere have been subducted beneath the Transhimalaya Batholith and the active continental margin of Asia, and the (Xigaze) forearc was gently folded only in Eocene times. This tectonic arrangement requires protracted collision between 55 and 40 Ma. This further implies that the Tsangpo Suture should be considered as a

614 double suture where remnants of the Tethys Ocean, the oceanic arc system with its back arc, and
615 the fore-arc of Eurasia have been subducted and/or imbricated in the orogen. These remnants may
616 contribute to the present-day thickness of the Transhimalaya.

617 Zircon and monazite ages track the generally in-sequence, southward progression of the orogenic
618 deformations at the expense of the Indian continent. The early collision is marked by drastically
619 slowed convergence between India and Eurasia, a change in sedimentation, and the beginning of
620 the prograde path of the UHP metamorphism. The Tethys oceanic lithosphere was pulling the
621 frontal parts of the India margin to great depth, occasionally recording UHP peak conditions at 55-
622 45 Ma. The early collision is also recorded in the Tethyan domain by early, mantle-derived
623 magmatism and a long magmatic and metamorphic history in the North Himalayan Domes, where
624 there is a lack of monazite data. Amphibolite facies metamorphics exhumed in these domes are
625 the witnesses of a thickened crust, whereas early plutons represent accreted material during
626 collision.

627 The precursors of the Himalayan Mountain Belt appear ca. 40 Ma. The Himalaya are an
628 intracontinental mountain belt at least 10 m.y. younger, as has been long recognized (e.g., Bird
629 1978; Chemenda et al. 2000).

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Figure captions

Fig. 1: Sketch map of the South Tibet-Himalayan orogenic belt with reference to the lithological and tectonic zones discussed in the text, adapted from several reviews, some referenced in the text.

Fig. 2: Simplified sections (located on Fig. 1) showing interpretation of crustal structures across the South Tibet-Himalayan orogenic belt. Magmatic ages discussed in text and reported in figures 3 and 4. Same colours as in figure 1. A: Tsangpo Suture - Central Himalaya section; deep reflectors (Moho and dashed thrust (?) lines from Hirn et al. (1984a), Hirn et al. (1984b), Nelson et al. (1996), Hauck et al. (1998) and Schulte-Pelkum et al. (2005). Deep within mantle seismogenic zone from De La Torre et al. (2007). B: Western Himalaya: Structural variations and Moho depth after Verma and Prasad (2009) and Li and Mashele (2009); Moho offset, crustal detachment and faults after Seeber et al. (1981), after Belousoff et al. (1980) beneath North-Kohistan and Karakoram.

Fig. 3: Histogram of crystallization ages of plutonic rocks defined by zircons. n = number of data. Several vaguely located samples did not permit minimize sampling bias. References used for this compilation are given in electronic supplement.

Fig. 4: Tectonic reconstruction of the India-Asia convergence system in late Cretaceous times. Options chosen are that (1) the main Indian units are inherited from extensional structures of the passive margin, including the MCT, which may have reactivated the breakaway fault

zone, and (2) intra-oceanic arcs existed all along the orogenic zone and were subducted beneath the Eurasian margin. Discussion in text.

Fig. 5: Histogram of $^{206}\text{Pb}/^{238}\text{U}$ zircon data for the North Himalaya granitoids. References in Supplementary material. Abbreviations as in text and figure 3.

Fig. 6: Simplified geologic and metamorphic cross section through the Himalaya of Central Nepal after Le Fort (1986). Ages from Corrie and Kohn (2011) and Kohn et al. (2004), orange = prograde monazite and blue = retrograde monazite. Intrusion age of Manaslu leucogranite from Harrison et al. (1999). Same colours as in Figure 1; section slightly to the west of, and parallel to the High Himalaya part of section A, Fig. 2.

Fig. 7: Histogram of $^{206}\text{Pb}/^{238}\text{U}$ monazite and zircon data are from the Everest region (Larson et al. 2011), Annapurna (Corrie and Kohn 2011) and Bhutan (Kellett et al. 2013; Regis et al. 2014; Rubatto et al. 2013). Other references in supplementary material.

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Fig. 5: Histogram of $^{206}\text{Pb}/^{238}\text{U}$ zircon data for the North Himalaya granitoids. References in Supplementary material. Abbreviations as in text and figure 3.

Fig. 6: Simplified geologic and metamorphic cross section through the Himalaya of Central Nepal after (Le Fort 1986). Ages from (Corrie and Kohn 2011) and (Kohn et al. 2004), orange = prograde monazite and blue = retrograde monazite. Intrusion age of Manaslu leucogranite from (Harrison et al. 1999). Same colours as in Figure 1; section slightly to the west of, and parallel to the High Himalaya part of section A, Fig. 2.

Fig. 7: Histogram of $^{206}\text{Pb}/^{238}\text{U}$ monazite and zircon data are from the Everest region (Larson et al. 2011), Annapurna (Corrie and Kohn 2011) and Bhutan (Kellett et al. 2013; Regis et al. 2014) and (Rubatto et al. 2013). Other references in supplementary material