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Invited review

The timing of India-Asia collision onset – Facts, theories, controversies

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ABSTRACT

The timing of initial collision between India and Asia has remained controversial for half a century. This paper attempts to review this crucial and hotly debated argument, describing first the different methods used to constrain the age of collision and discussing next the rationale, results, inferences and problems associated with each. We conclude that stratigraphy represents the best direct way to unravel collision chronology. Other methods focusing on the magmatic, metamorphic or paleomagnetic record provide additional fundamental constraints, but cannot provide a robust direct estimate of collision onset.

Initial collision in the central-eastern Himalaya is dated directly at the middle Paleocene (59 ± 1 Ma) by the abrupt change in sediment provenance recorded in trench settings. The quasi-synchronous unconformities documented along both Tethyan passive margin of India and active margin of Asia from Tibet to Zanskar-Ladakh confirm that orogeny was underway at the close of the Paleocene (56 Ma), well before the disappearance of marine seaways in the Himalaya during the early-middle Eocene (50–45 Ma). Sedimentary evolution and provenance changes in marine to fluvio-deltaic successions are recorded synchronously within error from the western to the central-eastern Himalaya, failing to provide conclusive evidence for diachronous collision.

These coherent observations are hard to reconcile with three widely cited hypotheses invoking either Paleogene arc-continent collision or Late Cretaceous ophiolite obduction, or the protracted existence of a Greater India Basin, which are all not favored after discussing the geological evidence critically point by point. A scenario no more complex than the one involving solely the passive continental margin of India and the active continental margin of Asia is needed to explain the geological evolution of the nascent Himalaya. The collision between the Tethys Himalaya and the Transhimalayan arc-trench system does represent the collision between India and Asia. Because the Yarlung Zangbo Ophiolite is the forearc basement of the Asian active margin, its obduction onto India could not have preceded the initial closure of Neo-Tethys. Ophiolite obduction began when collision began, in the middle Paleocene.

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“There is something fascinating about science. One gets such wholesale returns of conjecture out of such a trifling investment of fact.”

[Mark Twain, 1883, Life on the Mississippi, ch. 17]

1. Introduction

Continental collision, perhaps the most spectacular result of plate tectonics, generates the most elevated mountain belts on Earth and represents the main process to form large continental masses. Asia, the largest of continents, is composed of different blocks welded by collision events that took place at different stages through geological time. The most impressive, recent and still active of these is the collision between the Indian and Asian continental margins, which produced the Himalayas, the archetype of collision orogens and the key to interpret mountain ranges produced by continent-continent collision in the past (Gansser, 1964; Le Fort, 1975; Dewey et al., 1988; Yin and Harrison, 2000).

A precise determination of the timing and potential diachroneity of India-Asia collision onset is critical to an accurate calculation of crustal shortening, to unravel the subsequent tectonic evolution as well as the uplift mechanism of the Qinghai-Tibet Plateau, to understand the migration and evolution of faunas (Bossuyt and Milinkovitch, 2001) or to test the potential link between orogenic growth, global cooling and intensification of the Asian monsoon (Raymo and Ruddiman, 1992; An et al., 2001).

It is customary in the literature to assign a variety of ages to the India-Asia collision, ranging rather freely from as old as the Late Cretaceous to as young as the Oligo-Miocene (e.g., Blondeau et al., 1986; Garzanti et al., 1987; Yin and Harrison, 2000; Aitchison et al., 2007b; Wu et al., 2008; Huang et al., 2010; Najman et al., 2010; Hu et al., 2012; van Hinsbergen et al., 2012; Huang et al., 2015a). The confusion generated by such a wide range of ages has led many authors to conclude that collision might have taken place anytime between >65 and <40 Ma (e.g., Rowley, 1996; Zhu et al., 2005; Zhang et al., 2012). Researchers have thus been left to select in remarkable liberty the age that fitted best with their lines of reasoning. In the last decade, however, a growing body of evidence has contributed to constrain the age of collision with increasing accuracy and precision, through the study of stratigraphy, sedimentology, magmatism, metamorphism, structural geology and paleomagnetism. These new studies have left many critical recent findings unheralded, and motivates this review.

The aim of this article is to survey these crucial and controversial arguments, describing first the different methods used to constrain the onset of collision between India and Asia and discussing next the rationale, results, inferences and problems associated with each. Subsequently, we critically revisit several hypothetical scenarios proposed - including Paleogene arc-continent collision, Late Cretaceous ophiolite obduction, and the Cretaceous-Paleogene Greater Indian Basin - focusing on the geological evidence on which those ideas were based. Finally, we illustrate the timing of the India-Asia collision as presently understood, discuss the uncertainties concerning its diachroneity, and suggest future directions of geological research on the earliest stages of orogeny in the Himalayas and other mountain ranges.

1.1. Half century of inconclusive debate

Investigations on the timing of India-Asia collision onset can be traced back to the dawn of the plate-tectonic era. Powell and Conaghan (1973) and Chang and Zheng (1973) inferred a pre-middle Eocene collision onset based on the age of the youngest marine deposits found in the Himalayas and paleomagnetic evidence from the Indian Ocean. Molnar and Tapponnier (1975) interpreted a decrease in the rate of northward drift of India at ~40 Ma deduced from paleomagnetic data from the Indian Ocean floor as indicative of ongoing India-Asia collision. Molnar and Tapponnier (1977) added four other geological pieces of evidence to support their conclusion: (1) Upper Cretaceous "exotic blocks" found in the Indus-Yarlung suture zone (Gansser, 1964) indicate that the Tethyan Ocean still existed at 70 Ma; (2) Tethys Himalayan marine strata deposited throughout the Paleozoic and Mesozoic until the early Eocene suggest that collision initiated after the early Eocene; (3) the oldest mammals found spread from India to Mongolia are middle Eocene in age; (4) major mountain building in the Himalayas began in the Oligocene (<34 Ma).

Based on paleomagnetic evidence obtained by the Sino-French collaboration program in Tibet, Patriat and Achache (1984) reconstructed the drift histories of the Indian and Asian plates and proposed that India started to collide with Asia around 50 Ma. They suggested that collision with the Ladakh arc may have taken place slightly earlier around 54 Ma, but without major effect on India plate motion. Such collision diachroneity between the western and central Himalayas was widely adopted by subsequent researchers (e.g., Gansser, 1991; Rowley, 1996). In the Zanskar Range, Garzanti et al. (1987) recognized for the first time the onlap of Transhimalayan-derived, volcaniclastic fuviodeltaic sediments onto the Indian passive margin, corresponding to the end of marine sedimentation dated at the P8 foraminiferal biozone (middle of early Eocene), as well as a major disconformity between planktonic foraminiferal P5 and P6 biozones (Paleocene/Eocene boundary), thus constraining onset of India-Asia collision in the western Himalaya by ~56 Ma. Beck et al. (1995) found that accretionary-prism and trench strata in northwestern Pakistan were thrust onto India between 66 Ma and before 55.5 Ma, indicating collision in that time window. In the central Himalaya, Searle et al. (1987) preferred later collision onset at 50–40 Ma based on: (1) the change from marine to continental sedimentation in the Indus-Yarlung suture zone; (2) the end of Gangdese I-type granitoid intrusions; (3) Eocene S-type anatectic granites and migmatites in the Lhasa Block; (4) the beginning of compressional tectonics in the Himalayas. Rowley (1996) reviewed stratigraphic evidence from the Himalayas and concluded that only in the western Zanskar-Hazara region the record was well constrained, indicating that collision started in the late Ypresian (≤52 Ma). He further stated that east of Everest "*normal, shallow shelf-type carbonates extend into the Lutetian, without evidence of a change in sedimentation to the top of the section, so the start of collision must be still younger*", concluding that collision initiated in the late Ypresian in the west and progressed into the Lutetian in the east. Rowley (1998), based on subsidence analysis of the Zhepure Mountains succession in southern Tibet, supported such diachronous initiation of the India-Asia collision. Since then, the debate on diachronous versus quasi-synchronous collision became another issue of controversy (e.g., DeCelles et al., 2004; Ding et al., 2005; Najman et al., 2005). Yin and Harrison (2000) provided a comprehensive review including plate-kinematic, paleomagnetic, magmatic, stratigraphic and paleontological evidence, concluding that "*initial collision between India and Asia could have started as early as latest Cretaceous*

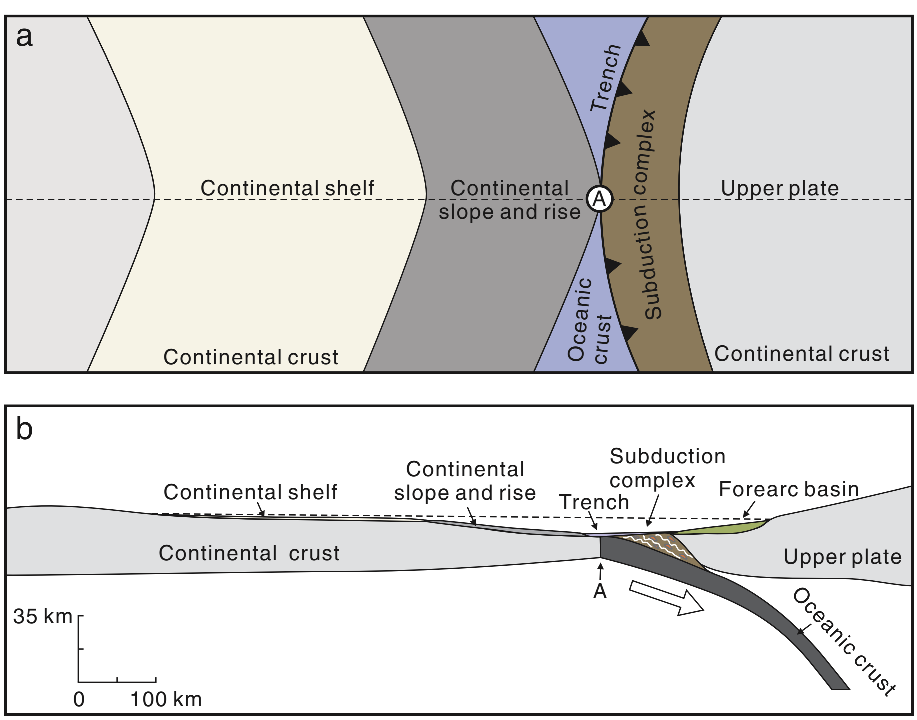

Fig. 1. The cartoon clarifies the definition of initial continental collision: A) plane view; B) profile.

time (~*70 Ma*)". Aitchison et al. (2000, 2007b; Abrajevitch et al., 2005) proposed that India collided first with an intraoceanic arc at ~55 Ma, and not with Asia until ~35 Ma. Based on paleomagnetic data and to account for a discrepancy between estimated convergence and shortening, van Hinsbergen et al. (2012) suggested that collision with Asia of a micro-continent now preserved as the Tethys Himalaya around 50 Ma was followed by progressive closure of a largely oceanic Greater India Basin "*along a subduction zone at the location of the Greater Himalaya*". In this scenario, the rest of India collided with Asia around 25–20 Ma. These controversial hypotheses have been discussed in numerous subsequent publications (e.g., Ding et al., 2005; Zhu et al., 2005; Najman et al., 2010; Yi et al., 2011; Hu et al., 2012; Lippert et al., 2014).

1.2. Definition of collision onset

In earlier times researchers used to define collision onset in rather vague terms, often equating initial collision with final closure of Neo-Tethyan seaways (e.g., Patriat and Achache, 1984; Blondeau et al., 1986; Searle et al., 1987). As a consequence, the end of marine sedimentation was used commonly as a proxy of collision onset (e.g., Rowley, 1996). Beck et al. (1995) were the first to define explicitly collision onset as the moment "*when the crust of the Neo-Tethys ocean was completely subducted at some point along the plate boundary*" (Fig. 1). This definition of initial continent-continent contact, equating with initiation of continental subduction and orogeny (in tectonic sense) is accepted here as in our previous studies (e.g., Hu et al., 2012, 2015a). Collision onset as defined by disappearance of oceanic lithosphere at one point is necessarily significantly older than the demise of each single leftover seaway connected with the disappearing Neo-Tethys and consequent complete cessation of marine sedimentation throughout the nascent Himalayan belt. Disappearance of oceanic lithosphere may have occurred at different times along the collision front because of irregular shape of either continental margin, potentially resulting in diachroneity along strike.

Mascle et al. (2012, p. 192) made a distinction between "*continental contact*" and "*continental collision*", and stated: "*Two continents come into contact with one another after the end of an oceanic subduction. Therefore, the thinned part of the continental margin, in this case, the Indian margin that was attached to the Tethys ocean, became subducted beneath the active margin, in this case, the South Asian margin. The continental collision occurred only later, when the contact zone is sufficiently thickened, giving rise to the first elevated continental reliefs by thrusting and, therefore, to the thickening of the upper crust of the subducting margin, in this case, the Indian crust*". There is no way to establish objectively at what point the contact zone would be "*sufficiently thickened*". Therefore we prefer our definition of continent contact as synonymous with continental collision onset.

2. Geologic outline of the Himalaya and southern Tibet

2.1. Tectonic units

The southern margin of Asia is represented by the Gangdese arc-trench system in the central and eastern Himalaya, and by the Kohistan-Ladakh arc in the western Himalaya.

The Indus-Yarlung suture zone (also named Indus suture zone by Molnar and Tapponnier, 1975; Himalayan suture by Gansser, 1980; Indus-Tsangpo suture by Allegre et al., 1984, Searle et al., 1987; Yarlung-Zangbo (Tsangpo) suture by Girardeau et al., 1984; Aitchison et al., 2000; IYSZ, Fig. 2), marked by discontinuous ophiolite complexes and serpentinite-matrix mélange, delineates the E-W trending contact between the Indian and Asian plates for over 2500 km (Gansser, 1980; Searle et al., 1987; Hébert et al., 2012; Dai et al., 2013; An et al., 2014, 2016).

Recently, in their overview of the studies on the Yarlung Zangbo Ophiolites, Wu et al. (2014b) suggested that ultramafic and mafic rocks were formed in separate stages and with no genetic relationships, based on the following observations: (1) mantle peridotites predominate over mafic rocks, cumulate gabbros and sheeted dykes being usually absent; (2) basalts directly overlie the peridotites and are significantly different in terms of formation age and Sr-Nd isotopic compositions (Liu et al., 2012; Dai et al., 2013). If this was the case, then the ultramafic rocks in the Yarlung Zangbo suture zone would be older than the basalts and formed in different tectonic settings. From the western to the eastern Himalaya, ultramafic rocks are cut by gabbros and dolerite dykes derived from depleted asthenospheric mantle and emplaced around 130–120 Ma (see reviews by Wu et al., 2014b; Zhang et al., 2016).

The Xigaze forearc basin lies north of the IYSZ as part of the southern active margin of the Asian plate. The basaltic unit of the ophiolites is stratigraphically overlain by a shallowing-upward megasequence including abyssal sediments at the base (Chongdui Formation), deep-sea turbidites (Ngamring Formation), and shelfal to deltaic facies at the top (Padana Formation) (Einsele et al., 1994; Dürr, 1996; Wang et al., 2012; An et al., 2014).

The Lhasa Block farther to the north is an E-W–trending tectonic domain subdivided into distinct northern, central, and southern terranes characterized by different sedimentary covers and magmatism (Zhu et al., 2011). In the southern Lhasa terrane, sedimentary strata are limited and mainly of Late Triassic–Cretaceous age (Zhu et al., 2013). Upper Paleozoic (meta)sedimentary and Upper Jurassic–Lower Cretaceous volcano-sedimentary strata are widespread in the central Lhasa terrane (Pan et al., 2004; Chung et al., 2005; Chu et al., 2006; Zhu et al., 2011). In the northern Lhasa terrane, sedimentary rocks are mainly Jurassic and Cretaceous and locally Triassic in age. Magmatic rocks emplaced in the southern and central Lhasa terranes from the Late Triassic to the Eocene (220–40 Ma) are characterized by distinct zircon U-Pb age spectra and Hf isotopic ratios (Ji et al., 2009; Zhu et al., 2011, 2013). Geochronological studies suggest that plutonic rocks in the southern Lhasa terrane formed in discrete magmatic events at 205–152 Ma, 109–80 Ma, 65–41 Ma, and 33–13 Ma (Chung et al., 2005; Zhu et al., 2011). The southern Lhasa terrane yields predominantly zircons with positive εHf(t) (Chu et al., 2006; Ji et al., 2009), whereas zircons with negative εHf(t) predominate in the central Lhasa terrane (Chu et al., 2006; Zhu et al., 2011). The different U-Pb age and εHf(t) signatures of the southern and central Lhasa terranes are of essence in sandstone provenance interpretation.

The remnants of the edge of the Indian continent are exposed south of the Indus-Yarlung suture, and traditionally subdivided into the Tethys, Greater, and Lesser Himalayas (Fig. 2). The Neoproterozoic to Eocene Tethys Himalayan Sequence represents the sedimentary succession of northernmost India (Sciunnach and Garzanti, 2012), including a southern and a northern zone (Ratschbacher et al., 1994; Fig. 2). The southern zone includes platform carbonates and diverse terrigenous units of Palaeozoic to Eocene age (Willems et al., 1996; Jadoul et al., 1998; Hu et al., 2012), whereas the northern zone is dominated by Mesozoic to Palaeocene outer shelf, continental slope and rise deposits (Li et al., 2005b). The western Himalaya succession is similar to that of southern Tibet in the upper part, starting from Carboniferous rifting of Neo-Tethys. The lower part includes >2.5 km thick, Neoproterozoic to Lower Palaeozoic shallow-marine siliciclastic and carbonate deposits overlain with angular unconformity by Ordovician "molassic" conglomerates (Garzanti et al., 1986).

South of the Tethys Himalaya lies the Greater Himalaya, comprising medium to high-grade metasediments and Cambro-Ordovician granitoid protoliths, and structurally defined by bounding faults: the South Tibet Detachment Zone above and the Main Central Thrust below (Burchfiel and Royden, 1985; Webb et al., 2011). Below the Main Central Thrust, the Lesser Himalaya includes Proterozoic to Eocene strata (DeCelles et al., 1998), which experienced up to lower-amphibolite-facies metamorphism in the Main Central Thrust footwall (Caddick et al., 2007; Célérier et al., 2009). The famed inverted metamorphism of

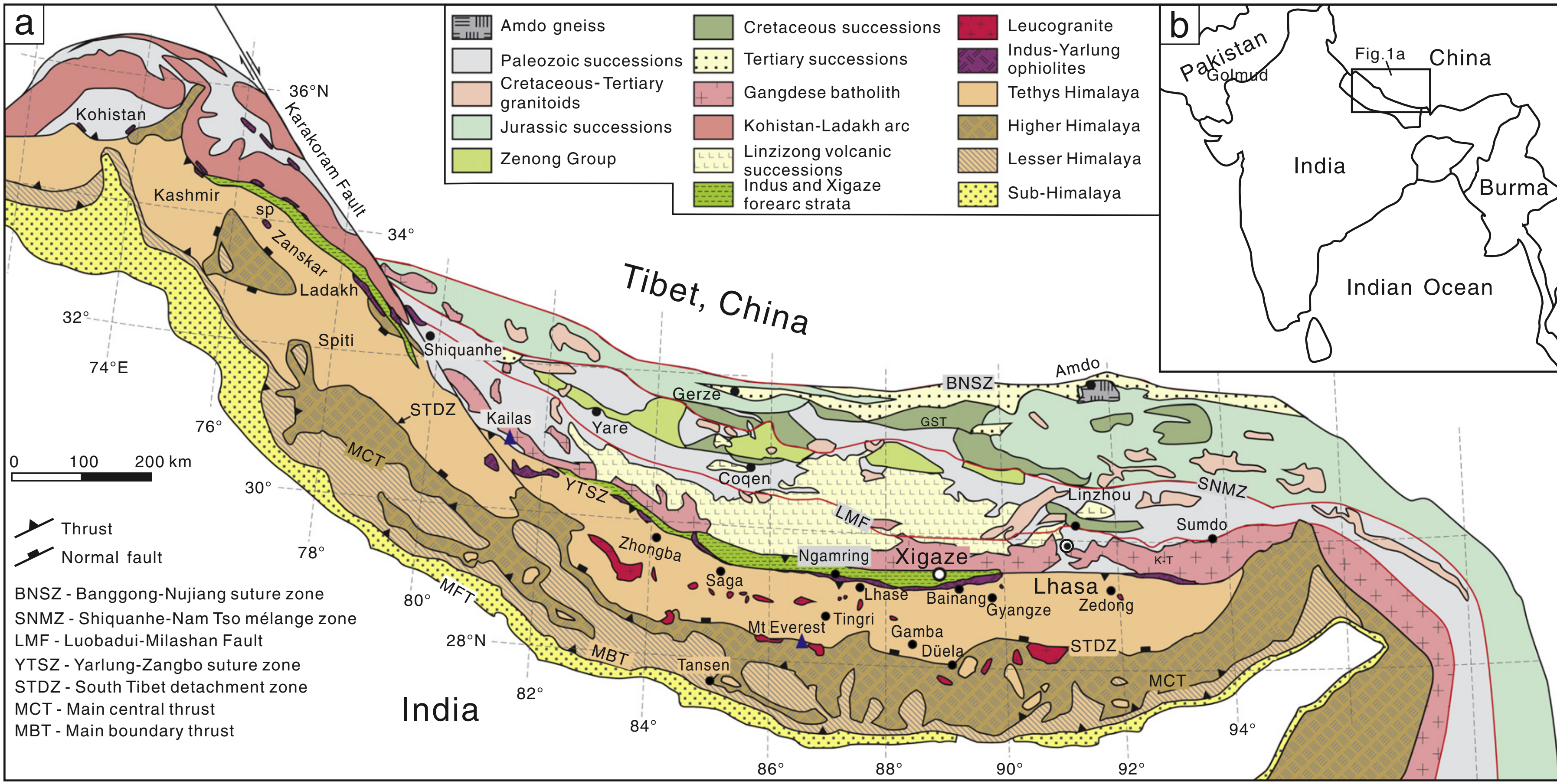

Fig. 2. Simplified geological map of the Himalaya and southern Tibet.
(Revised from Pan et al., 2004).

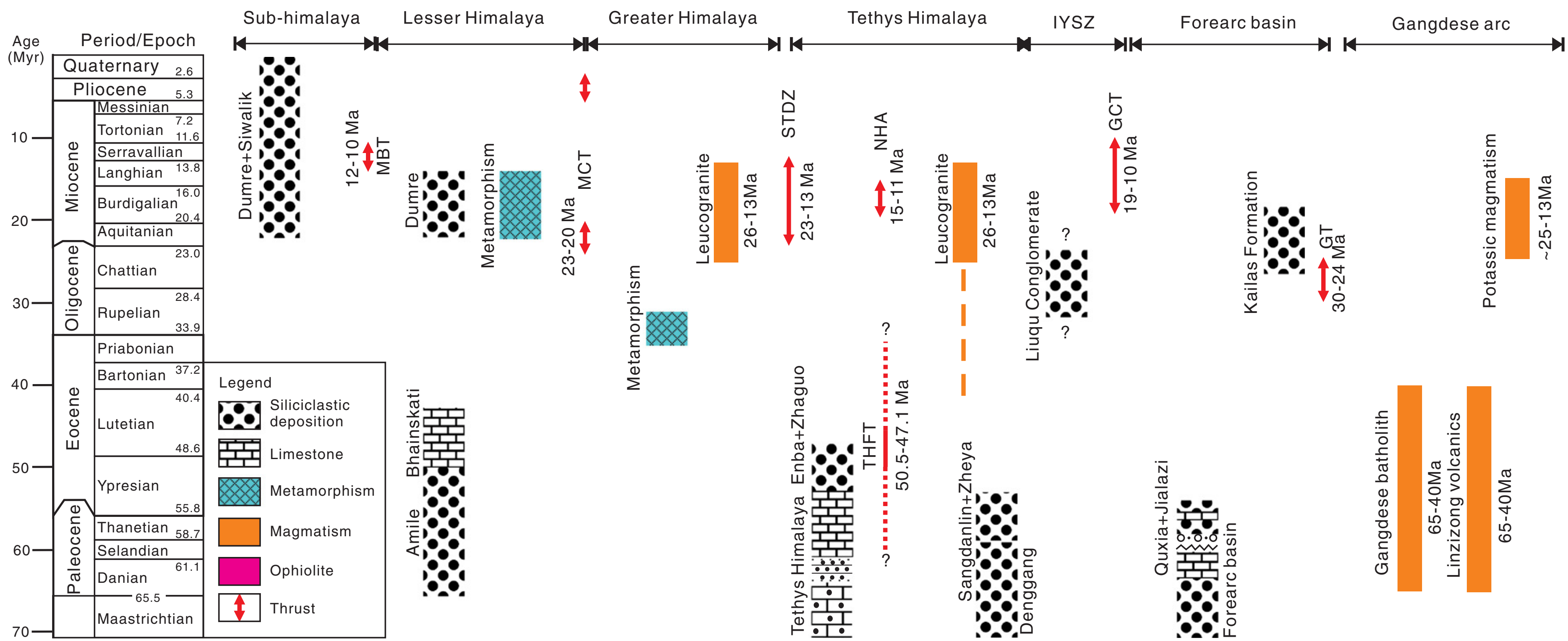

Fig. 3. Summary of the ages of major faults and of magmatic and metamorphic events in the central Himalaya. Latest marine deposits include the Bhainskati Formation in Nepal, the Zongpu, Enba and Zhaguo formations in the southern Tethys Himalaya, the Sangdanlin and Zheya formations in the northern Tethyan Himalaya, and the Quxia and Jialazi formations from the southern margin of the Asian plate. Sources include Chung et al. (2005), Yin (2006), Najman (2006) and others mentioned in the text. MBT: Main Boundary Thrust; MCT: Main Central Thrust; STDZ: South Tibet Detachment Zone; THFT: Tethys Himalaya Fold-Thrusts; NHA: North Himalayan Antiform (=North Himalayan Gneiss Domes); GCT: Greater Counter Thrust; IYSZ: Indus-Yarlung suture zone; GT: Gangdese Thrust.

the Himalaya is displayed by the decreasing metamorphic conditions recorded from the upper parts of the Greater Himalaya down to the middle portions of the Lesser Himalaya (e.g., Medlicott, 1864; Le Fort, 1975; He et al., 2015). Cenozoic foreland-basin strata are preserved on top of the Lesser Himalaya in Nepal (DeCelles et al., 1998, 2004; Najman et al., 2005) and in NW India (Najman and Garzanti, 2000; Ravikant et al., 2011) (Fig. 2). The Bengal and Indus Fans developed on remnant-ocean floor since the Eocene, following the onset of the India-Asia collision (Fig. 2; e.g. Najman et al., 2008; Zhuang et al., 2015).

The principal sedimentary, metamorphic, magmatic and tectonic events in different Himalayan domains are depicted in Fig. 3.

2.2. Upper Cretaceous-Paleogene stratigraphy

Cretaceous through Paleogene rocks within and surrounding the Indus–Yarlung suture zone provide the critical record of collision timing. In this subsection we shall briefly describe the main features of the Upper Cretaceous to Eocene sedimentary succession in southern Tibet. For a general description of correlative units in the central-western Himalaya from Nepal to NW India the reader is referred to Gaetani and Garzanti (1991), Garzanti (1999), Sciunnach and Garzanti (2012).

Throughout the article, correlation between the relative biostratigraphic scales and absolute ages are according to Gradstein et al. (2012).

2.2.1. Southern Tethys Himalaya in Tibet

In the Tingri-Gamba area of southern Tibet (Fig. 2), the upper Albian–Coniacian (100–86 Ma) Gambacunkou Formation (Fig. 4) consists of monotonous, ~600 m thick grey calcareous marlstones and marly limestones (Willems et al., 1996). The overlying Coniacian to Santonian (86–83 Ma) Jiubao Formation is ~80 m-thick and dominated by regularly bedded limestones (Wendler et al., 2009). Both units include abundant planktonic foraminifera and were deposited in a pelagic environment at depths of a few hundred meters (Willems et al., 1996; Hu et al., 2012).

The contact between the Jiubao and the overlying Zhepure Shanpo Formation is a paraconformity with most of the Campanian missing (Fig. 5A; Hu et al., 2012). The Zhepure Shanpo Formation contains at the base calcareous layers yielding reworked planktonic foraminifera of latest Campanian to early Maastrichtian age (72–66 Ma) (Hu et al., 2012). This 190 m-thick shallowing-upward sequence includes marly limestones and marls overlain by up to very-coarse-grained channelized quartzarenites (Willems et al., 1996) representing hyperpycnal flows or storm-surge turbidites with wave-reworked tops. Next, interbedded marls and sandstones are followed by another quartzarenite interval overlain in turn by fossiliferous marls dated at the Danian (Wan et al., 2002).

The overlying Danian (66–62 Ma) Jidula Formation consists of quartzose coastal sandstones interbedded with mudrocks and sandy limestones yielding gastropods, ostracods and foraminifers, ranging in thickness from 75 m at Tingri to 180 m at Gamba. Felsitic volcanic rock fragments, a few weathered feldspars and poor heavy-mineral suites including Cr-spinel indicate provenance from cratonic India in the south, rejuvenated during the Deccan volcanic event (Garzanti and Hu, 2015). The top of the unit is overlain with sharp transgressive boundary (Fig. 5B) by the shallow-marine carbonates of the 350–440 m-thick Zongpu Formation, yielding in the lower part foraminifers of late Danian age (~62 Ma; Wan et al., 2002). The unit consists of thick-bedded and locally dolomitized limestones, followed by nodular limestones and by thick-bedded biocalcarenites (Willems et al., 1996) extremely rich in larger benthic foraminifers indicating a late Danian to early Eocene age (Ypresian, SBZ 10–12, 50–48 Ma; BouDagher-Fadel et al., 2015). The unit, accumulated on a carbonate ramp, records a deepening trend overall, interrupted at Gamba by an unconformity corresponding to the Palaeocene/Eocene boundary and marked by a conglomerate with intraformational limestone clasts (Fig. 5C; Zhang et al., 2012; Li et al., 2015b).

The Enba Formation (~32 m in Gamba, ~105 m in Tingri) disconformably overlying the Zongpu Formation (Fig. 5D) consists of greenish-grey calcareous mudrocks intercalated with storm-deposited oolitic and bioclastic limestones in the lower part and with litho-quartzose sandstones in the upper part, documenting the final stages of marine deposition in southern Tibet. Above, the Zhaguo Formation (>20 m in Gamba, ~75 m in Tingri; Fig. 5E) consists of red mudrocks with intercalated intraformational caliche conglomerates and lenticular sandstones with scoured base, becoming more numerous, thicker and coarser-grained up-section testifying to the progradation of a subaerial deltaic plain. Both Enba and Zhaguo sandstones are litho-feldspatho-quartzose, containing abundant felsic to intermediate volcanic rock fragments and subordinate sedimentary rock fragments. Petrographic composition, detrital zircon geochronology and geochemistry testify to a drastic change in provenance, documenting the first arrival of Transhimalayan volcanic arc detritus from the north to the southern Tethys Himalayan passive continental margin (Zhu et al., 2005; Cai et al., 2008; Najman et al., 2010; Hu et al., 2012; Li et al., 2015b). The youngest zircon grains in the Enba (54 ± 1 Ma at Gamba, 54–52 Ma at Tingri) and Zhaguo formations (53 ± 1 Ma at Gamba, ~52 Ma at Tingri but for one grain yielding 43 Ma) constrain the maximum depositional age of these units as late Ypresian (Hu et al., 2012; Li et al., 2015a). The reported occurrence of the calcareous nannofossil *Sphenolithus pseudoradians*, however, has suggested that the preserved top of the succession may be as young as the Priabonian (NP 20, ~35 Ma; Wang et al., 2002; Jiang et al., 2016).

2.2.2. Northern Tethys Himalaya in Tibet

The ~700 m thick Sangdanlin section is exposed south of the Indus-Yarlung suture zone. The Denggang Formation (~110 m) includes coarse-grained turbiditic quartzarenites and a quartzo-lithic basalticlastic bed interbedded with green shales, both derived from India, containing zircons yielding Early Cretaceous U-Pb ages (Wang et al., 2011; Wu et al., 2014a; DeCelles et al., 2014), and capped by greenish chert (units 10 and 11 in Wang et al., 2011) (Fig. 6).

In the conformably overlying Sangdanlin Formation (90 m; Fig. 5F), red cherty mudrocks in the lower part yielded radiolaria of middle Paleogene (Selandian) radiolarian zones RP 4–6 (RP code numbers refer to Sanfilippo and Nigrini, 1998; Hu et al., 2015a) (Fig. 6). The upper part of the Sangdanlin Formation includes Indian-derived quartzose turbidites interbedded at several intervals with feldspatho-litho-quartzose volcaniclastic turbidites with subordinate chert, shale/siltstone and phyllite grains derived from the Gangdese and central Lhasa domains (DeCelles et al., 2014; Wu et al., 2014a; Hu et al., 2015a). Intercalated red and green cherty mudrocks yielded radiolaria of Selandian age (RP6; ~60 Ma; Hu et al., 2015a) (Fig. 6). The youngest U-Pb ages of detrital zircons are ~58 Ma throughout the unit (Wu et al., 2014a; DeCelles et al., 2014; Hu et al., 2015a). The overlying Zheya Formation (~500 m) chiefly consists of feldspatho-litho-quartzose volcaniclastic turbidites interbedded with poorly fossiliferous grey mudrocks and cherts. Exotic blocks of conglomerate, sandstone and rudistid-bearing carbonate are embedded within silty shales in the upper part; conglomerate blocks contain chert, serpentinite, quartzarenite, siltstone and carbonate clasts.

The Upper Cretaceous succession exposed in the Gyangze area consists of the Aptian to lower Santonian Gyabula and upper Santonian to lower Campanian Chuangde formations (Li et al., 2005b), characterized by grey and red mudrocks deposited in continental-slope and basinal environments, respectively. Slumped blocks of pelagic limestones occur in the Chuangde Formation, which is overlain by the chaotic Zongzhuo Formation (Li et al., 2005b). This mèlange unit of the Zongzhuo Formation with black shale matrix includes up to pluridecametric blocks of chert yielding radiolaria of Late Jurassic-Paleocene age (Liu and Aitchison, 2002), thus indicating deposition not earlier than the Paleocene. Blocks of reddish microbialitic limestone, basalt, conglomerate, and quartz-

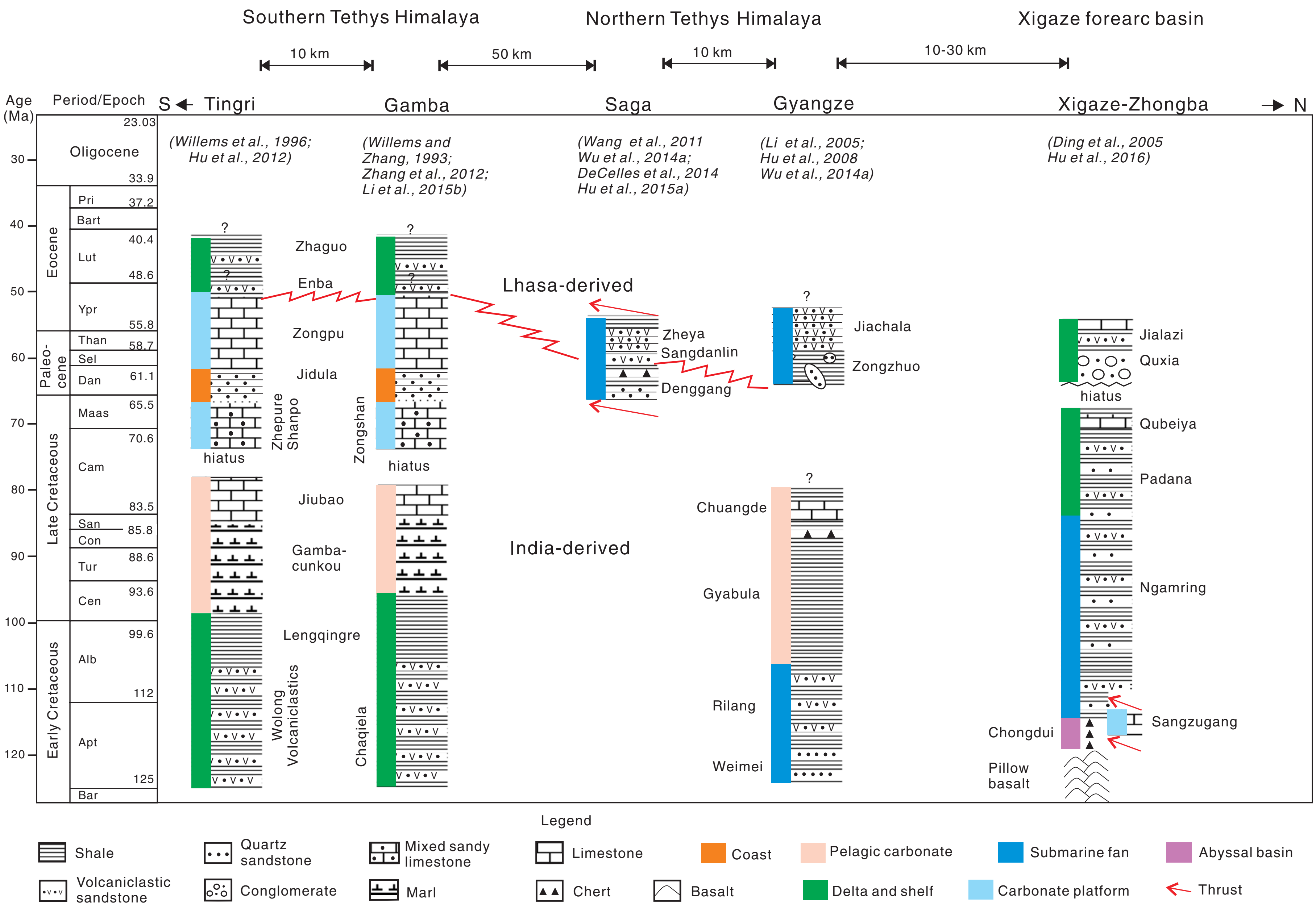

Fig. 4. Stratigraphic columns of Cretaceous–Paleogene strata in the Himalayan orogen. Correlation between biostratigraphic scales and absolute ages is according to Gradstein et al. (2012).

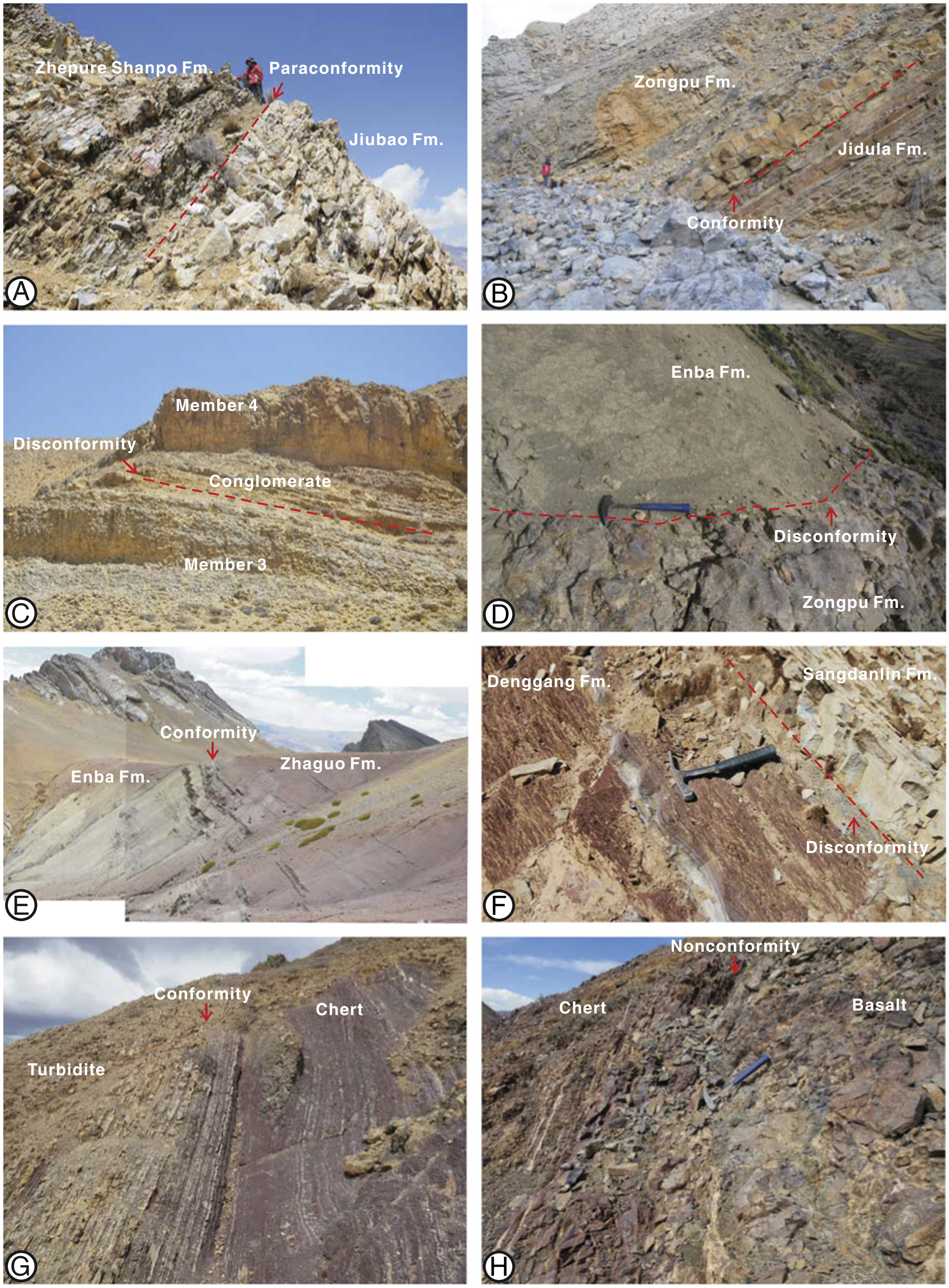

Fig. 5. Representative field photographs showing stratigraphic contacts mentioned in the text. (A) Paraconformity between Zhepure Shanpo and Jiubao formations, Gelamu section, Tingri (GPS: N28°42′47.9″, E86°45′13.4″); (B) paraconformity between Jidula and Zhongpu formations, Shenkezha section, Tingri (GPS: N28°41′33.6″ E86°42′50.3″); (C) disconformity between Member 3 limestones and conglomerates of Member 4 of the Zongpu Formation, Zengbudong section, Gamba (GPS: N28°16′52.9″, E 88°32′17.1″); (D) disconformity between Zongpu and Enba formations, Qumiba section, Tingri (GPS: N28°41′37.4″, E86°43′23.4″); (E) conformity between Enba and Zhaguo formations, Zhajia section, Tingri (GPS: N 28°41′12.9″, E 86°43′58.6″); (F) disconformity between Denggang and Sangdanlin formations, Sangdanlin section, Saga (GPS: N29°15′28″, E85°14′52″); (G), conformity between Chongdui chert and overlying turbidities; (H) undisputable stratigraphic contact between the cherts of the Chongdui Formation and the underlying basalt of the Yarlung-Tsangpo Ophiolite, Qunrang section, Xigaze (GPS: N29°19′17.7″, E89°02′40.2″).

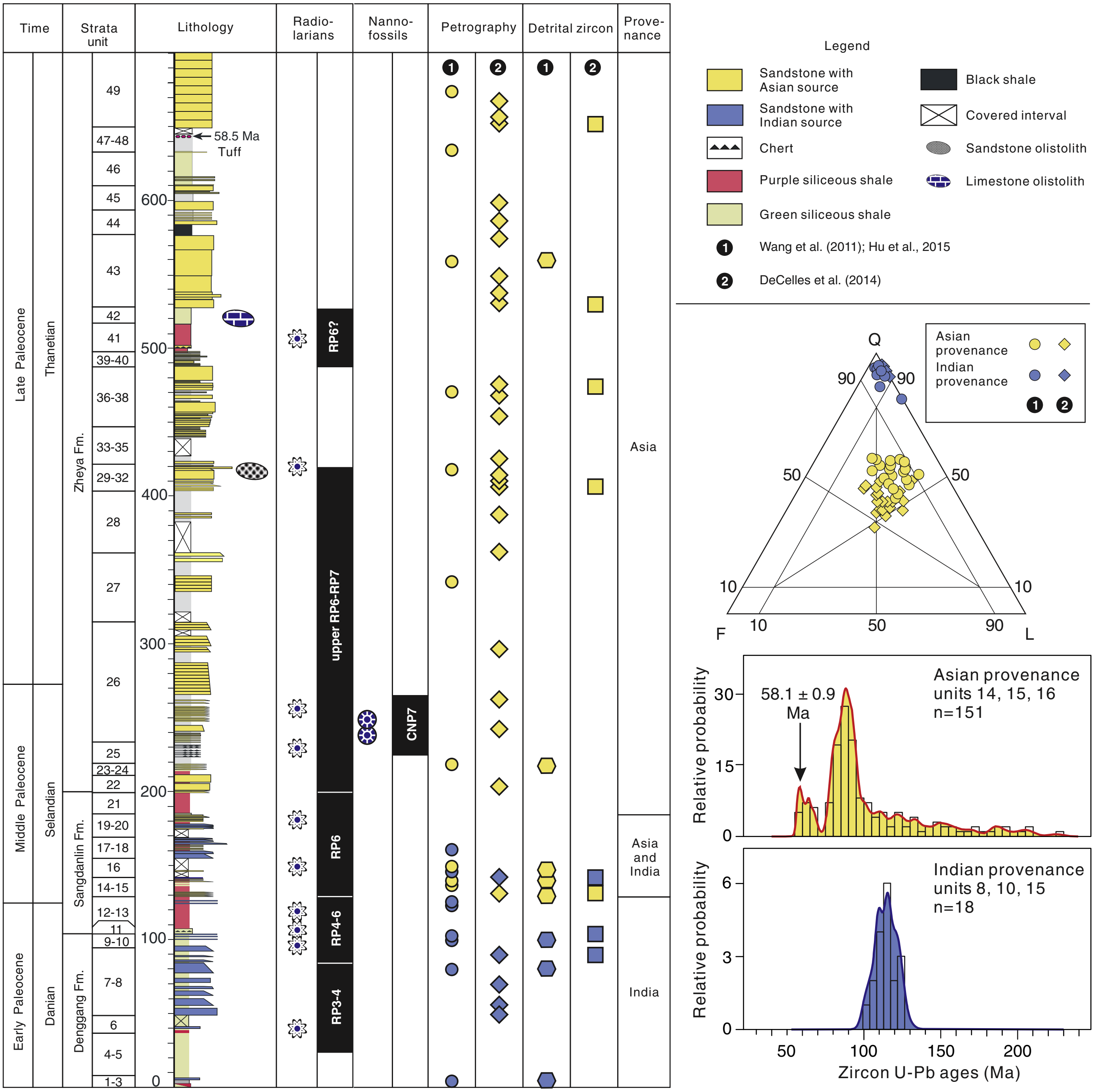

Fig. 6. Integrated bio- and chrono-stratigraphy, sandstone petrography and detrital zircon U-Pb ages of the Sangdanlin section. The distribution of radiolaria and calcareous-nannofossils, together with the youngest U-Pb ages of detrital zircons constrain the age of interbedded Indian- and Asian-derived turbidites (units 14–16) within the middle Paleocene (Selandian RP6 and CNP7 biozones, respectively). QFL triangular diagrams (Q-Quartz; F-Feldspar; L-Lithics) and detrital zircon U-Pb age patterns highlight the sharp compositional difference between Indian-derived quartzose turbidites and Asian-derived volcano-plutonic sandstones.
(Modified after Hu et al., 2015a).

poor volcaniclastic to sedimentaclastic sandstone resembling Xigaze forearc-basin turbidites also occur. Detrital zircons from sandstone blocks of the Zongzhuo Formation yielded U-Pb ages mainly between 129 and 77 Ma, suggesting provenance from the Gangdese arc, but also as young as 56 Ma (Cai et al., 2008; our own unpublished data). The overlying Jiachala Formation consists of turbiditic sandstones and interbedded mudrocks from which dinoflagellate and pollen assemblages of Paleocene-early Eocene age have been reported (Li et al., 2005a).

2.2.3. Xigaze forearc basin

The Xigaze forearc-basin succession is subdivided into the Chongdui, Sangzugang, Ngamring, Padana and Qubeiya formations (e.g. Wang et al., 1999, 2012). The overlying Quxia and Jialazi formations represent the youngest deposits within the basin (Ding et al., 2005; Hu et al., 2016).

The Chongdui Formation is subdivided into two members: (1) 73 to 97 m-thick, purplish-greenish radiolarian chert interbedded with siliceous mudstone deposited at abyssal depths on pillow basalts (Fig. 5H; An et al., 2014); (2) 200 m-thick, dark-grey, thin-bedded sandstones and mudrocks with minor limestones at the top (Fig. 5G; Wu, 1984; Ziabrev et al., 2003). The lower chert member contains a late Barremian–late Aptian (127–115 Ma) radiolarian association (Ziabrev et al., 2003), whereas the overlying sandstone member contains detrital zircons with a youngest age peak of 116 Ma (late Aptian; Wu et al., 2010; Dai et al., 2015).

The 60–230-m-thick Sangzugang Formation, exposed in the northern part of the basin in tectonic contact with the Ngamring Formation, consists of dark-grey, thick-bedded or massive bioclastic limestones with abundant benthic foraminifera, rudists, corals, and minor bivalves. It was deposited in reefal carbonate lagoon environments along the northern side of the forearc basin during the late Aptian (116.5–113 Ma; TLK1c biozone of Boudagher-Fadel et al., 2016).

The Ngamring Formation, conformably overlying the Chongdui Formation, represents the main turbiditic fill of the Xigaze forearc basin, with thickness between 1 and 4 km. Based on stratigraphic and structural analysis, Wang et al. (1999, 2012) identified five megasequences, including channelized conglomerates, turbiditic feldspatho-lithic to feldspatho-quartzo-lithic volcaniclastic sandstones, and mudrocks with minor limestones and marls deposited in submarine-fan to slope environments. Both the underlying contact with the Chongdui Formation and the overlying contact with the Padana Formation are stratigraphic and conformable. Ammonites and planktonic foraminiferal associations indicate a late Aptian to late Coniacian age for the Ngamring Formation (120–86 Ma, Wiedmann and Dürr, 1995; Wan et al., 1998).

The final filling stage of the Xigaze forearc basin is represented by the shallow-marine to fluvio-deltaic Padana and Qubeiya formations of latest Cretaceous age (An et al., 2014). The 733–2000 m thick Padana Formation consists of varicolored feldspatho-quartzo-lithic volcaniclastic sandstones interbedded with mudrocks, minor conglomerates and limestones deposited in shelf and prodelta settings, passing upward to delta-front, delta-plain and finally subaerial delta-plain environments (An et al., 2014). The youngest detrital zircon U-Pb ages constrain the unit to the Santonian–early Campanian (86–75 Ma; An et al., 2014).

The Qubeiya Formation (~200 m thick in the Zhongba area; Hu et al., 2016) conformably overlies the Padana Formation and consists of yellowish-grey sandy wackestones intercalated with fine-grained sandstones. Abundant fossils include larger benthic foraminifera (mainly *Lepidorbitoides* spp.), bivalves, ammonites, gastropods and crinoids, indicating deposition on the inner shelf during the Maastrichtian (70–66 Ma; Liu et al., 1988; Hu et al., 2016).

The overlying Quxia and Jialazi formations represent syn-collisional continental deposits (Hu et al., 2016). The ~105 m-thick Quxia Formation starts with unfossiliferous grey to reddish floodplain mudrocks intercalated with coarse-grained quartzo-litho-feldspathic volcaniclastic sandstones and followed by alluvial-fan conglomerates. The nature and geometry of the stratigraphic contact between the Qubeiya Formation and the overlying Quxia Formation remains uncertain because of poor outcrop conditions. An angular unconformity was suggested by Ding et al. (2005). Volcanic and volcaniclastic pebbles and cobbles prevail over granitoid, chert, limestone and sandstone clasts. The youngest obtained U-Pb zircon age is 65.5 ± 1.0 Ma (Hu et al., 2016).

The ~240 m thick Jialazi Formation, conformably overlying the Quxia Formation, includes sandy limestones and feldspatho-litho-quartzose volcaniclastic sandstones deposited in a distal fan-delta environment. In the lower part of the unit, rich faunas include benthic and few planktonic foraminifera indicating the Thanetian (SBZ 4; late P4; 58–57 Ma). Two tuff layers yielded weighted means for the youngest U-Pb age peak of 54.9 Ma ± 0.7 Ma and 55.7 Ma ± 0.5 Ma. In the upper part of the unit, diversified benthic foraminifera indicate an earliest Eocene age (SBZ 5; 56–55 Ma). The youngest detrital-zircon U-Pb age is 54.0 Ma. Cr-spinels in both Quxia and Jialazi formations have similar geochemistry as those in the underlying Xigaze forearc strata (Hu et al., 2014, 2016).

In the Kailas area of southwestern Tibet, the earliest Eocene Dajin Formation (56–54 Ma; Yan et al., 2006; Wang et al., 2015) comprises massive or normally graded, matrix- or clast-supported conglomerates, massive or cross-laminated sandstones, and poorly laminated mudstones indicating deposition from subacqueous to subaerial debris flows. Abundant sandstone and mudstone clasts indicate extensive recycling of forearc strata ultimately derived from the Gangdese magmatic arc. Deposition of the Dajin Formation is interpreted to document penecontemporaneous fold-thrust deformation in the Gangdese forearc (Wang et al., 2015).

3. Methods used to constrain collision onset

3.1. Stratigraphy and sedimentology

The sedimentary record provides a direct way to effectively pinpoint the timing of collision onset and to reconstruct the salient phases of the subsequent syn-collisional tectonic evolution (Garzanti et al., 1987; Rowley, 1996; Najman, 2006). As scrutinized below, several distinct arguments can be used to that goal, including provenance changes revealing tectonic changes in the source areas, the cessation of marine deposition and transition to continental sedimentation, and/or the development of unconformities and abrupt changes in sedimentary environments or basin subsidence.

3.1.1. Provenance changes

3.1.1.1. Rationale. After initial collision, detritus from the upper plate (i.e., Asia) can be transported and deposited on top of the lower plate (i.e., India). It is undisputable that the moment when detritus from the Asian margin reached and was deposited on top of the Indian continental margin, collision was well underway. Asian-derived detritus would reach first the distal toe of the Indian continental margin at abyssal depths, and at later times would downlap progressively southward on the shallower-water inner parts of the lower-plate margin. Terrestrial sediments fed from Asia and carried all the way by rivers onto India mark the final stage in which all depositional space between the two continents was filled completely. Collision onset can thus be dated directly only from the sedimentary record of the Indian continental rise (Fig. 1), whereas diachronous progradation of Asian material onto the Indian shelf only provides younger and younger minimum ages for collision onset.

As DeCelles et al. (2014) stated, "*as two continents approach during an impending collision, their respective detrital aprons will comingle. Although turbidite fans constructed of continental detritus may extend much more than 1000 km offshore, they generally are deposited upon oceanic crust at abyssal depths. Thus, when the detritus of one continental landmass is deposited upon the continental shelf or slope of another continental landmass, the two opposing continents are engaged in initial collision. The inherent topographic and bathymetric asymmetry of the impending collisional landscape, with a subduction zone and relatively high elevation magmatic arc along the margin of the upper plate and a marine passive margin along the leading edge of the lower plate, places the detrital record of initial approach and collision onto the lower plate*".

All provenance methods, including detrital-zircon geochronology and isotope geochemistry can be employed to detect the expected major compositional change recorded in the lower-plate succession, from typically quartz-rich sandstones of the pre-collisional passive-margin stage to lithic-rich sandstones of the syn-collisional stage. The timing of such provenance change can be dated by a combination of biostratigraphic, magnetostratigraphic, and zircon chronostratigraphic techniques (Hu et al., 2015a).

3.1.1.2. State of the art. In southern Tibet, the first arrival of Asian detritus onto the Indian margin is recorded in deep-water Paleocene strata of the northern Tethys Himalaya (Sangdanlin, Zongzhuo and Jiachala formations; Ding et al., 2005; Wang et al., 2011; Wu et al., 2014a; DeCelles et al., 2014), and next in shallow-water lower Eocene strata of the southern Tethys Himalaya (Enba Formation; Wang et al., 2002; Zhu et al., 2005; Najman et al., 2010) (Fig. 4). Such a major provenance change has long been reported from shallow-marine to fluvio-deltaic lower Eocene strata of the northwestern Himalaya (Kong and Chulung La formations; Garzanti et al., 1987), northern Pakistan (Patala and Balakot formations; Critelli and Garzanti, 1994 p.271; Garzanti et al., 1996), and northern India (Subathu Formation; Najman and Garzanti, 2000).

The Zongzhuo Formation exposed in the Gyangze area was held to be the oldest sedimentary unit to receive detritus from Asia (Cai et al., 2011; Wu et al., 2014a). The Zongzhuo Formation is however a mélange ("wildflysch with exotic blocks"; Tapponnier et al., 1981; Burg and Chen, 1984) lying unconformably on folded and cleaved basinal Mesozoic sequences of the Indian margin (Burg and Chen, 1984). The Jiachala Formation was also fed from Asia and, although it has been reported to yield Palaeocene/early Eocene dinoflagellates (Li et al., 2005a), the youngest detrital zircons it contains are not younger than around 77 Ma. Considering that magmatism was continuous throughout the Late Cretaceous-Paleogene in the Gangdese source area, detrital-zircon chronostratigraphy suggests a Campanian (Late Cretaceous) depositional age. More detailed biostratigraphic work is needed on both Zongzhuo and Jiachala formations to constrain their depositional ages firmly.

The best Palaeocene stratigraphic record of the distal Indian continental rise is preserved in the Sangdanlin section of the Saga area (southern Tibet), where quartzose sandstones sourced from India are intercalated with, and progressively replaced upward by, feldspatho-litho-quartzose volcanoplutoniclastic sandstones derived from Asia (Wang et al., 2011; DeCelles et al., 2014; Wu et al., 2014a) (Fig. 6). After an initial dispute about the age of such provenance change, ascribed to either the Palaeocene (Ding, 2003) or the early Eocene (Li et al., 2007; Wang et al., 2011), Wu et al. (2014a) and DeCelles et al. (2014) demonstrated that the youngest age cluster of Asian-derived zircons in the Sangdanlin Formation is not younger than 60–58 Ma. Furthermore, DeCelles et al. (2014) dated a tuff bed from the upper part of the overlying Zheya, ~510 m above the first arrival of Asian-derived detritus, as 58.5 ± 0.6 Ma. Finally, Hu et al. (2015a) coupled radiolarian and nannofossil biostratigraphy with detrital-zircon geochronology from the Sangdanlin section to constrain robustly the time when Asian-derived detritus was first deposited onto the distal edge of India as 59 ± 1 Ma (Selandian, middle Paleocene).

3.1.1.3. Conclusion and suggestions. The detailed biostratigraphic and provenance analysis of the crucial Sangdanlin section, documenting the most distal, deep-marine stratigraphic record of the Indian lower-plate margin, allowed to place a firm direct constraint on the timing of collision onset at 59 ± 1 Ma. In shallow-marine inner margin successions (e.g., Enba-Zhaguo and Kong-Chulung La formations), the arrival of Asian-derived detritus was delayed to 5–10 Ma later.

Coupling stratigraphic and provenance techniques represents an effective approach to date the onset of collision between two continents directly. The accuracy of the estimate however depends on the location and preservation of the sections chosen for analysis, because only in the most distal lower-plate-margin successions the provenance change from lower plate (Indian) to upper plate (Asian) sources can reveal the correct timing of collision onset. More proximal successions of the Indian continental margin, such as those exposed in Tingri, Gamba or the Zanskar Range, can only provide a minimum age for collision onset by provenance analysis.

3.1.2. Cessation of marine sedimentation

3.1.2.1. Rationale. At the onset of collision, oceanic lithosphere disappears at one point and the two continental margins come into direct contact (Fig. 1). Deep-marine conditions, however, are bound to remain for a considerable period of time before the trough between the shelf edge on one side and the trench-slope break on the opposite side is filled by sediments and/or eliminated by tectonic uplift and deformation (Fig. 1). And even after that, shallow seas may persist for a further amount of time before the nascent orogen grows in width and the regions on both sides of the suture are uplifted entirely well above sea level. The cessation of marine sedimentation thus can by no means be considered as contemporaneous with collision onset, but must follow it by several million years at least. Through detailed biostratigraphic and geochronological studies of the youngest marine strata from both lower and upper plates we can only obtain a minimum age falling considerably short of the exact timing of the first continent-continent contact.

3.1.2.2. State of the art. The youngest marine deposits in southern Tibet are represented by the Enba Formation in the southern Tethys Himalaya, by the Zheya and Jiachala formations in the Saga and Gyangze areas of the northern Tethys Himalaya, or by the Jialazi and Dajin formations in the Zhongba and Kailas areas of the Asian (Transhimalayan) margin.

The Enba Formation was deposited in shallow-marine and prodelta environments. The age assigned to this unit varies much, from the mid-Ypresian (Blondeau et al., 1986; nannofossil zone NP12, 52.8–50.6 Ma: Najman et al., 2010), to the Lutetian (Willems et al., 1996), and even to as young as the late Priabonian in the Tingri area (nannofossil zone NP20, 34 Ma: Wang et al., 2002), or to the early Priabonian in the Gamba-Düela area (~35 Ma: Wan, 1987; Li and Wan, 2003; Jiang et al., 2016). The Zheya Formation consists of deep-water turbidities deposited at late Selandian to early Thanetian times as constrained by combined radiolarian and nannofossil biostratigraphy and U-Pb zircon-dating of an interbedded tuff (DeCelles et al., 2014; Hu et al., 2015a). Deep-water deposits of the Jiachala Formation were ascribed to the Paleocene-early Eocene (Li et al., 2005a). The shallow-marine Jialazi and Dajin formations reach the Ypresian (~54 Ma, Wang et al., 2015; BouDagher-Fadel et al., 2015; Hu et al., 2016).

In the northwestern Himalaya, marine sedimentation in the Indus forearc basin persisted until the late Ypresian (SBZ 11, 51–49 Ma; see Henderson et al., 2010). The youngest marine sediments of the Indian passive margin are lower Ypresian in the inner shelf (~54 Ma; SBZ 6–8 for the top of the Dibling Formation: Nicora et al., 1987; late SBZ 7/early SBZ 8: Najman et al., 2016) and upper Ypresian in the outer shelf (53–51 Ma for the Kong Formation; zones P8 and SBZ 9–10; Garzanti et al., 1987; Green et al., 2008; Mathur et al., 2009). The youngest U-Pb ages of detrital zircons are 56.3 Ma for the Kong Formation and of 53.7 Ma for the overlying deltaic Chulung La Formation (Najman et al., 2016).

In the Himalayan foreland basin south of the proto-Himalayan range, marine deposition continued well into the Lutetian (43–41 Ma; SBZ 4–9 to SBZ 14–16 for the Subathu Formation of northern India: Bhatia et al., 2013; SBZ 13–14 for the Bhainskati Formation of Nepal: Matsumaru and Sakai, 1989).

3.1.2.3. Conclusion and suggestions. Marine deposition continued well into the Eocene on both sides of the suture zone, along the southern margin of Asia as along the northern margin of India, thus much later than Neo-Tethyan oceanic lithosphere had disappeared at the point of initial continent-continent contact. Sea arms can and do persist for several Myr to even tens of Myr after oceanic lithosphere has been consumed at one point (Sinclair, 1997), which implies that cessation of marine facies only provides a minimum age for initial collision, often far younger than collision onset. Modern examples include the seaway between Australia and New Guinea, the Taiwan Strait, or the Persian Gulf between Arabia and the Zagros orogen. Equating the disappearance of marine deposits with collision onset (Searle et al., 1987; Rowley, 1996; Aitchison et al., 2007b) is thus incorrect.

3.1.3. Intermontane sedimentation along the suture zone

3.1.3.1. Rationale. Alluvial fan to fluvio-deltaic sandstones and conglomerates receiving detritus from one or both of the uplifting collided margins may be deposited in terrestrial environments on either side of the suture zone at different stages of the syn-collisional evolution of the nascent orogenic belt. To provide a robust time constraint, strata must demonstrably overlie stratigraphically both collided continents, or contain detrital materials derived from both continents. Because of the lack of age-diagnostic marine fossils, the age of terrestrial strata is generally hard to establish with biostratigraphic methods, and can only be constrained by detrital-zircon chronostratigraphy if quasi-continuous magmatism in one of the source areas can be assumed.

3.1.3.2. State of the art. Two distinct suites of coarse subaerial clastic units are identified in the Indus-Yarlung suture zone of southern Tibet: (1) the Kailas Formation known by a variety of local stratigraphic names (Qiuwu, Dazhuka, Luobusha, Gangrinboche, Gansser, 1964; Yin et al., 1999; Aitchison et al., 2002; DeCelles et al., 2011; Wang et al., 2013) and exposed between the Xigaze forearc and the Gangdese arc; (2) the Liuqu Conglomerate exposed adjacent to the Yarlung-Zangbo ophiolites. The Kailas Formation was deposited in alluvial fans, braided to meandering rivers and lakes (e.g. DeCelles et al., 2011; Wang et al., 2013) during the late Oligocene–Miocene, as indicated by sporopollens and U-Pb age of interbedded tuffs (~26–18 Ma; Aitchison et al., 2009; Li et al., 2010; DeCelles et al., 2011; Wang et al., 2013; Carrapa et al., 2014). Recently, Leary et al. (2016a) documented that deposition of the Kailas Formation occurred at 26–24 Ma in western Tibet (81°E), at 25–23 Ma north of Lazi (87.8°E), at 23–22 Ma near Dazhuka (89.8°E), and as late as 18 Ma southwest of Lhasa (92°E), indicating that basin formation propagated eastward at a rate of approximately 300 mm/y. The Kailas Formation in the Kailas area was mainly fed from the Gangdese magmatic arc to the north, whereas during its latest stages it was sourced from the Tethyan Himalayan thrust belt in the hanging wall of the Great Counter thrust to the south (DeCelles et al., 2011). Similarly, in the Xigaze area the Qiuwu Formation was derived entirely from the Gangdese magmatic arc, whereas the overlying Dazhuka Formation contains clasts derived from the Gangdese arc, the Xigaze forearc basin, and the Yarlung Zangbo suture (Wang et al., 2013; Leary et al., 2016a).

The Liuqu Conglomerate, deposited in alluvial fans and braided rivers (Davis et al., 2002), was assigned a mid-late Eocene age on paleobotanical evidence (Tao, 1988; Fang et al., 2004), but a late Eocene to Oligocene age based on sporopollens (Wei et al., 2009). Low-temperature thermochronology favors an even younger, latest Oligocene-Early Miocene age (Li et al., 2015b). Leary et al. (2016b) suggested for the Liuqu Conglomerate an age as young as ~20 Ma. Detrital sources for the Liuqu Conglomerate include the Yarlung Zangbo ophiolitic suture and units of Indian affinity. Gravel derived from the Gangdese arc massif, such as granite or intermediate-felsic volcanic rocks, was not detected (Davis et al., 2002; Wang et al., 2010a,b). However, Wang et al. (2010a,b) found pebbles of lithic-rich sandstones interpreted to be recycled from Xigaze forearc-basin strata (see Section 4.3.3 below). Recently, Leary et al. (2016b) suggest that the sediments of the Liuqu Conglomerate were transported north-northwest from the hanging wall of a coeval thrust fault system where Indian-sourced sediments, ophiolites and accretionary mélanges outcropped. They interpreted that Asian zircons found in the Liuqu Conglomerate were recycled northward after being incorporated into accretionary mélanges along the southern Asian margin prior to India-Asia collision.

3.1.3.3. Conclusion and suggestions. The sedimentation of coarse clastic units took place at different times along the suture zone, and even several tens of Myr later than collision onset. Their age may be much younger than the youngest marine deposit, thus providing an even less reliably usable criterion to assess the timing of collision onset. Such terrestrial clastic wedges resulted from distinct episodes of tectonic uplift and/or subsidence associated with successive phases of orogenic growth, and thus provide crucial information on the syn-collisional to post-collisional evolution of the belt, rather than on the date of its birth.

3.1.4. Unconformities and abrupt sedimentary changes

3.1.4.1. Rationale. Regional unconformities and abrupt changes of sedimentary environments may be recorded at different times on both continental margins, either before, during, or after initial continent-continent collision.

3.1.4.2. State of the art. In the southern Tethys Himalaya, unconformities associated with abrupt changes of depositional paleoenvironment are documented at four stratigraphic positions between Late Cretaceous and Paleogene times. These are: 1) the unconformity between the Jiubao Formation and the overlying Zhepure Shanpo Formation (Fig. 5A), spanning the latest Santonian to late Campanian (83–72 Ma, Wu et al., 2011) and corresponding to a change from deep-water pelagic to continental-shelf sedimentation (Willems et al., 1996; Hu et al., 2012); 2) the depositional boundary between the Jidula and Zongpu formations (Fig. 5B), dated at the late Danian (~62 Ma, Wan et al., 2002) and marking a change from beach to inner carbonate-ramp settings; 3) the Paleocene-Eocene unconformity (Fig. 5C) occurring in the upper part of the Zongpu Formation, marked by a conglomerate bed with intraformational carbonate clasts and corresponding to an abrupt change from middle carbonate-ramp to restricted-lagoonal facies (Wang et al., 2010b; Hu et al., 2012; Li et al., 2015b); 4) the early Eocene unconformity between the Zongpu and Enba formations, marked by a hardground (Fig. 5F) with iron oxides and representing an abrupt change from carbonate ramp to prodelta environments (Hu et al., 2012; Li et al., 2015b; Boudagher-Fadel et al., 2015).

On the Asian margin, the Upper Cretaceous Qubeiya Formation, deposited on a shallow inner shelf until the latest Cretaceous (~66 Ma), is overlain unconformably by the Paleocene-Eocene Quxia and Jialazi formations (Ding et al., 2005; Hu et al., 2016). The latter units were deposited in a fan-delta between 58 and 54 Ma after a hiatus of ~8 Ma (Hu et al., 2016), and document a major change in sporopollen and foraminiferal assemblages (Wan et al., 2002; Li et al., 2008).

All of these unconformities have been claimed to correspond with collision onset. Yin and Harrison (2000) hypothesized early collision around 70 Ma also based on the sedimentary change documented in the Tethys Himalaya by Willems et al. (1996). Hu et al. (2012) proposed that the entire Zongpu Formation might have been deposited as a carbonate ramp within an underfilled foreland basin, thus taking its base around 62 Ma as the possible timing of collision onset. Zhang et al. (2012) and Li et al. (2015b), as earlier on Garzanti et al. (1987) in the northwestern Himalaya, interpreted the Paleocene-Eocene unconformity as related to the passage of the orogenic wave associated with ongoing collision. And even the unconformity between the Zongpu and Enba formations was interpreted as the sedimentary response of the Indian margin to collision onset (Zhu et al., 2005; Najman et al., 2010). Finally, Hu et al. (2016) interpreted the unconformity between the Qubeiya and Quxia formations as the sedimentary response of the Asian margin to collision onset, thus favoring collision between 66 and 58 Ma.

3.1.4.3. Conclusion and suggestions. All abrupt sedimentary changes recorded by the Indian margin succession around 70 Ma, 62 Ma, 56 Ma and 50 Ma, or by the Asian margin succession between 66 and 58 Ma, may or may not be related to collision onset, because alternative explanations can be indicated for all of them, including tectonic events other than collision onset, magmatic flare-ups, or sea level changes. It is widely known that unconformities may develop in passive margins or forearc basins without any need to invoke collision events (e.g., McNeill et al., 2000; Ando and Tomosugi, 2005).

3.1.5. Tectonic character of sedimentary basins

3.1.5.1. Rationale. The onset of continent-continent collision theoretically determines a change in tectonic style of sedimentary basins on both sides of the suture zone. On the lower plate, passive-margin sedimentation ends, and syn-collisional foreland-basin sedimentation begins. On the upper plate, forearc-basin sedimentation ends as well, generally replaced by deposition in syncollisional basins. The stratigraphic expression of such a change is however difficult to define and detect operationally.

According to theory (e.g., Crampton and Allen, 1995; DeCelles and Giles, 1996; DeCelles, 2012), following initial underthrusting of the lower plate beneath the upper plate, a flexural bulge should develop within 0.1 Myr of initial collision and migrate progressively forelandward. The passage of such orogenic wave would produce the uplift of a topographic high, resulting in an unconformity that can be used to monitor ongoing syn-collisional deformation of the lower plate. As discussed above for facies and provenance changes, also tectonic changes migrate in time, and are thus recorded earlier by distal successions at the edge of the lower-plate margin and progressively later on by proximal successions deposited at a distance from the suture zone.

3.1.5.2. State of the art. On the Indian margin, tectonic uplift - interpreted as related to the passage of an orogenic wave originating from the point of initial continent-continent contact - is documented by the major disconformity recorded around the Paleocene/Eocene boundary all along the southern Tethys Himalaya from the Zanskar Range in the west (Garzanti et al., 1987) to southern Tibet in the east (Li et al., 2015b). According to this interpretation, the uppermost part of the Zongpu Formation with the overlying Enba and Zhaguo formations in southern Tibet, and the equivalent Kesi, Kong and Chulung La formations in the Zanskar Range, were deposited in a syn-collisional basin (Hu et al., 2012; Sciunnach and Garzanti, 2012).

On the Asian margin, the transition from forearc-basin to syn-collisional sedimentation is documented by the unconformity separating the Qubeiya and Quxia formations (Ding et al., 2005; Hu et al., 2016). The Dajin Formation, containing conglomerates recycling Gangdese magmatic rocks and older sedimentary rocks, was deposited in a wedge-top basin based on the assumption that collision began prior to 58 Ma (Wang et al., 2015), as discussed above. In the western Himalaya, final transition to syncollisional sedimentation is documented on the Asian margin by lower Eocene fan delta deposits (Nummulitic Limestone) passing upward to alluvial fan deposits (Nurla Formation; Garzanti and Van Haver, 1988; Henderson et al., 2010)

3.1.5.3. Conclusion and suggestions. As we presently understand the tectonic evolution in the earliest stages of collision, the first continent-continent contact took place at 59 ± 1 Ma, as proved by the robustly dated provenance change in the Sangdanlin Formation of the distal Indian margin (Hu et al., 2015a). Comingling turbidite fans derived from India and Asia marked the transition from passive-margin to trench sedimentation, continuing with the Zheya Formation above (DeCelles et al., 2014). In the proximal Indian margin, the tectonic effects of collision were felt only 3 ± 1 Ma later, as documented by the Paleocene/Eocene unconformity identified all along the southern Tethys Himalaya from the Zanskar Range to southern Tibet. Such a paleotectonic reconstruction is coherent and plausible, but it is based on the stratigraphic and provenance data illustrated in previous paragraphs above rather than on independent evidence. In other words, our capacity of establishing the pre-collisional or syn-collisional nature of a sedimentary basin depends on the robustness of our stratigraphic observations. We should firmly avoid to interpret the latter according to our preferences for the former, not to find ourselves caught in the pitfall of circular reasoning.

3.2. Paleomagnetic constraints

Paleomagnetism is one of the principal techniques for reconstructing continental drift and plate tectonics. Paleomagnetic investigations of the India-Asia collision started in the 1980s and early 1990s (e.g., Zhu et al., 1981; Pozzi et al., 1982; Achache et al., 1984; Klootwijk, 1984; Patriat and Achache, 1984; Patzelt et al., 1996; Appel et al., 1998) and were improved recently by using more sophisticated analytical procedures and larger datasets.

3.2.1. Paleolatitude overlap

3.2.1.1. Rationale. The timing of collision between India and Asia can be constrained by defining when the paleomagnetically determined paleolatitudes from the continental margins on both sides of the Indus-Yarlung suture started to overlap. This requires accurate and continuous record of the paleolatitude history of the Lhasa Block and Tethys Himalaya, which in principle can be determined by the apparent polar wander paths (APWPs) of Eurasia and India (Torsvik et al., 2012). In practice, post-collisional upper-crustal shortening within both Asia and the Himalaya limits our ability to predict paleolatitudes from APWPs, and robust paleomagnetic poles must be obtained directly for both Lhasa Block and Tethys Himalaya to constrain collision chronology.

3.2.1.2. State of the art. In the northern margin of India, paleomagnetic investigations have been hindered by pervasive remagnetization of Tethys Himalayan strata (Appel et al., 2012). Natural remanent magnetizations (NRMs) with a primary origin have been reported only from a few Cretaceous to Paleogene units (Klootwijk and Bingham, 1980; Patzelt et al., 1996; Yi et al., 2011; Huang et al., 2015b; Yang et al., 2015a; Ma et al., 2016) (Table DR1). Paleolatitudes of the Tethys Himalaya of 54.8° ± 3°S (reference position at IYSZ: 29°N, 88°E, same for below) and 48.2° ± 5.7°S were determined, respectively, from the Lower Cretaceous (138–130 Ma) Wölong volcaniclastic sandstones in the Tingri area and Lakang lavas (134–131 Ma) in the Cuona area (Huang et al., 2015b; Yang et al., 2015a). A similar paleolatitude of 45.8° ± 6.1°S was recently defined from the Lower Cretaceous Sangxiu lavas (135–124 Ma) in the Ngagyze area (Ma et al., 2016). Lower Aptian volcaniclastic sandstones of the Thakkhola region in Nepal yielded a paleolatitude of 44.4° ± 3.7°S (Klootwijk and Bingham, 1980). The only reliable paleomagnetic pole for the Late Cretaceous (68 ± 3 Ma) was obtained from limestones at Gamba and Duela, giving a paleolatitude of 5° ± 3.5°S (Patzelt et al., 1996; Dupont-Nivet et al., 2010). Primary remanence has also been separated from Paleogene limestones in the same areas, giving paleolatitudes of 6.3° ± 3.5°N for the lower Zongpu Formation (60 ± 2 Ma) and 10.8° ± 2.5°N for the upper Zongpu Formation (57 ± 2 Ma) (Patzelt et al., 1996; Yi et al., 2011). These paleomagnetic data would suggest a much larger extent of "Greater India" in the Late Cretaceous and early Paleogene than in the Early Cretaceous. Such a discrepancy cannot be ascribed to counterclockwise rotation of India since the Early Cretaceous (Huang et al., 2015a), and was used to support the Greater India Basin hypothesis (van Hinsbergen et al., 2012).

For the Lhasa Block, using the paleomagnetic-data filtering procedure of van Hinsbergen et al. (2012) and the maximum likelihood method of Arason and Levi (2010), a paleolatitude of 15.6° ± 2.2°N at ~120 Ma (Table DR1) was re-calculated after data of Sun et al. (2010), Chen et al. (2012), Ma et al. (2014), Li et al. (2016) and Yang et al. (2015b), with a dispersion of the virtual geomagnetic poles (DVGPs) of 15.1° consistent with the expected value for time-averaging secular variation at ~15°N (Johnson et al., 2008). Another study of the Xigaze forearc sedimentary rocks (maximum depositional age of ~129 Ma) yielded a similar paleolatitude of 16.1° ± 4.8°N after correction for inclination shallowing (Huang et al., 2015c). A higher paleolatitude of 24.1° ± 7.3°N was obtained from basaltic andesites erupted at 110–100 Ma (Chen et al., 2016), although with a DVGPs of 12.6° lower than expected at this latitude, indicating that paleosecular variation may have been undersampled.

The Late Cretaceous paleolatitudes of the Lhasa Block determined from both volcanic and sedimentary rocks are controversial. The Upper Cretaceous red beds of the Shexing Formation in Maxiang (100–72 Ma), the Takena Formation in Penbo (105–90 Ma) and the Qushenla Formation (100–83 Ma) in northeastern Gerze yielded paleolatitudes of 23.8° ± 2.3°N, 23.7° ± 2.2°N and 23.9° ± 3.2°N, respectively (Table DR1; Tan et al., 2010; Sun et al., 2012; van Hinsbergen et al., 2012; Chen et al., 2016) after correction for inclination shallowing by the E/I method (Tauxe and Kent, 2004). The Upper Cretaceous Daxiong red beds in the Coqen area (98–92 Ma; Sun et al., 2015) gave a lower paleolatitude of 17.5° ± 2.4°N after applying E/I correction (personal communication by Tianshui Yang; Tauxe and Kent, 2004; Yang et al., 2015b). A much lower paleolatitude of 9.4° ± 2.6°N was calculated for Upper Cretaceous volcanic rocks (92–80 Ma) in the Coqen, Yare and Shiquanhe basins (Tang et al., 2013; Yi et al., 2015) using the method of Arason and Levi (2010). Paleogene latitude estimates for the Lhasa Block were obtained from the regionally extensive Linzizong Group, including the Dianzhong, Nianbo and Pana formations (Achache et al., 1984; Chen et al., 2010, 2014; Dupont-Nivet et al., 2010; Liebke et al., 2010; Sun et al., 2010; Tan et al., 2010). Recent studies showed that only volcanic rocks (54–52 Ma) and volcaniclastic sandstones (52 ± 1 Ma) of the Pana Formation in the Linzhou basin retain a primary remanence, yielding paleolatitudes of 18.6° ± 3.8°N (DVGPs: 14.8°, explained by secular variation) and 18.8° ± 2.7°N (after correction for inclination shallowing) (Table DR1; Chen et al., 2010, 2014; Dupont-Nivet et al., 2010; Tan et al., 2010; Huang et al., 2013, 2015d). Similar paleolatitudes were retrieved from the partially remagnetized Linzizong Group (52 ± 0.6 Ma) in the Nanmulin basin (Huang et al., 2015e) (Table DR1). Latest Eocene (~35 Ma) paleolatitudes of 22.7° ± 3.1°N and 24.7° ± 3.8°N for the Lhasa Block were determined, respectively, from Kangtuo red beds in the Gerze basin after correction for inclination shallowing (Ding et al., 2015) and from lava flows in the southern Qiangtang Block (Lippert et al., 2011) (Table DR1).

3.2.1.3. Conclusion and suggestions. Using a reference point located at 29° N, 88°E on the IYSZ, paleolatitudes for the northern Tethys Himalaya and southern Lhasa Block are calculated to have started to overlap at the latitude of ~19°N at 52 ± 4 Ma (95% confidence intervals; see Fig. 7). This age should be considered as a minimum estimate for initial collision, which had to take place a few million years earlier if a typical forearc dimension and the potential crustal shortening of the Tethys Himalayan sequence north of Gamba and Duela are added to the leading edge of the Asian and Tethys Himalayan margins (Lippert et al., 2014).

Although imprecise because of large error bars, paleolatitude overlap represents a valuable independent method to constrain the age of collision onset.

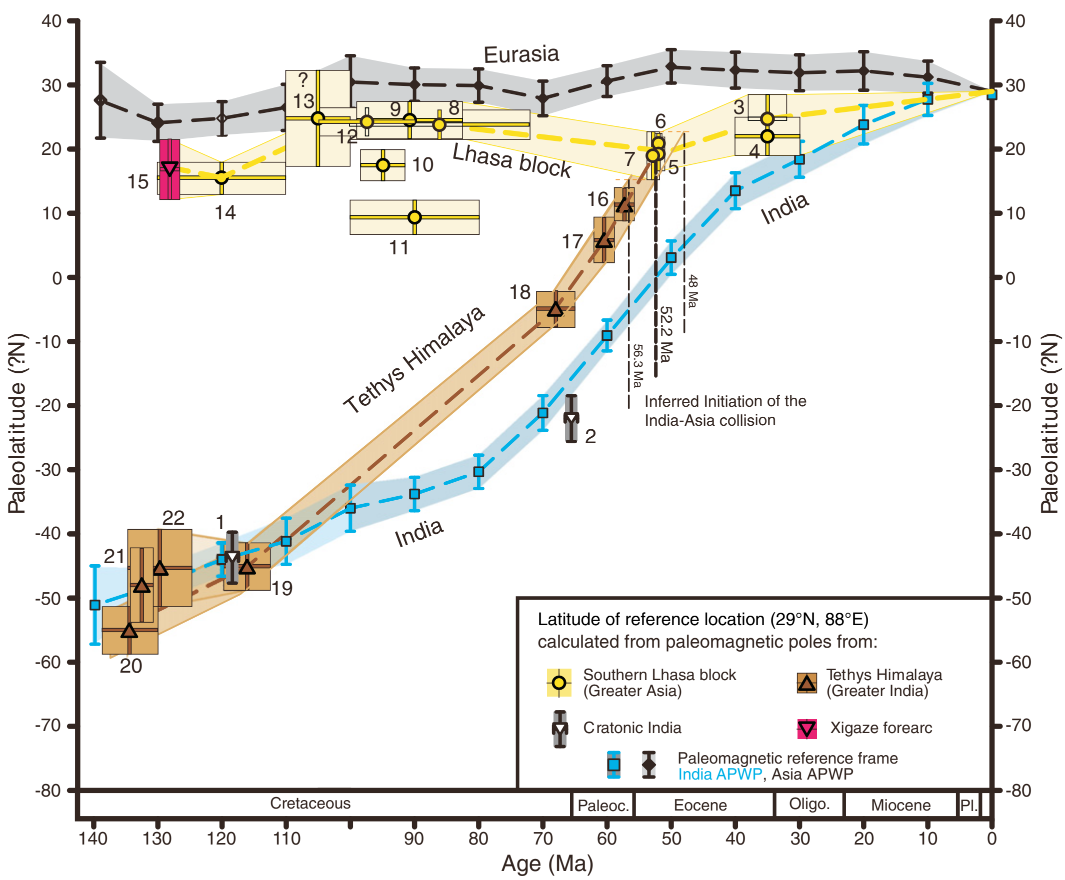

Fig. 7. Paleolatitudes of a reference site (29°N, 88°E) located on the present-day position of the Indus–Yarlung suture zone in Eurasian, Greater Asian, Tethys Himalayan, and Indian reference frames, with horizontal and vertical bars representing error bars for the age and paleolatitiude, respectively. Numbers correspond to paleomagnetic poles described and listed in Table DR1. Pole 13 may undersample the paleosecular variation. The paleomagnetically determined collision age is 52.2 ± 4 Ma (95% confidence interval); if forearc and crustal shortening are considered, initial collision should have taken place a few Myr earlier. Note that paleolatitude of Tethys Himalaya overlapped with that of India in the Early Cretaceous, but not in the Late Cretaceous to early Paleogene.

3.2.2. Convergence rate

3.2.2.1. Rationale. Continental crust is thicker and less dense than oceanic crust. When oceanic subduction ends and continental subduction begins, increasing buoyancy and resistance are expected to cause a notable deceleration of plate convergence. The change in convergence rate between India and Eurasia as reconstructed from the pattern of magnetic anomalies in the Indian Ocean was therefore widely used to constrain the timing of the India-Asia collision.

3.2.2.2. State of the art. Molnar and Tapponnier (1975) first found that the relative motion between India and Asia dropped from 100–180 mm/yr in the Late Cretaceous and Paleogene to ~50 mm/yr after ~38 Ma, and concluded that India first touched Asia during the Eocene. More detailed data subsequently indicated that this deceleration may have been caused instead by a decrease in the spreading rate of the Eastern Indian Ridge near anomaly 22 time (~50 Ma) (e.g., Johnson et al., 1976). Others (e.g., Liu et al., 1983) have correlated the India-Asia collision with the change in relative motion between India and Africa observed at anomaly 20 time (~44 Ma). Patriat and Achache (1984) suggested that the sudden change of spreading rate followed by irregularities in the motion of India observed at the time of anomalies 22 (49.7 Ma) and 21 (47.7 Ma) were caused by the closure of Neo-Tethys and onset of the India-Asia collision. Paleomagnetic results from the Ninety-east Ridge indicated a distinct reduction in the motion of India at ~55 Ma, interpreted as the completion of suturing with Asia initiated earlier (Klootwijk et al., 1992). More recently, van Hinsbergen et al. (2011a) concluded that the sharp deceleration in the motion of India observed at ~55–50 Ma most likely resulted from increased resistance to subduction and decreased slab pull after collision onset as well as from restrengthening of lithosphere-asthenosphere coupling upon demise of the Deccan plume. The revised plate reconstructions by White and Lister (2012) indicates that the Indian plate motion was distinctly episodic through time, with multiple accelerations (at 85–64 Ma, 62–61 Ma, 55–52 Ma, 47–45 Ma, 20–19 Ma, 18–16 Ma, 14–13 Ma and 11–10 Ma) and decelerations (at 60–58 Ma, 52–51 Ma, 48.5–47 Ma, 45–39.5 Ma, 26.5–25 Ma, 19–18.7 Ma, 15.5–14.5 Ma, 12–11 Ma and 9.8–5.5 Ma). The authors argued that the timing of each deceleration might indicate a separate accretion event during the northward progression of India, and that independent geological data are needed to reconstruct the complex tectonic evolution of the Himalayan orogen.

3.2.2.3. Conclusion and suggestions. Although the deceleration of India-Asia convergence documented at ~55–50 Ma was widely believed to be collision-related, the main problem of constraining collision onset by this criterion is that geodynamic explanations for deceleration are far from unique, and include the demise of the Deccan plume effect, strain accumulation within the plate circuit, or other major accretion events. Moreover, Zhu et al. (2015) argued that slowdown of India may have resulted from break-off of the subducting Neo-Tethyan oceanic slab rather than from the India-Asia collision. To identify the effect of each potential factor requires not only higher resolution data but independent geological observations as well.

3.2.3. Remagnetization

3.2.3.1. Rationale. Remagnetization (i.e., the acquisition of a geologically stable remanence long after the formation of the rock) is an insidious problem when paleomagnetism is applied to paleogeographic reconstructions (e.g., van der Voo and Torsvik, 2012). Widespread remagnetization events are often linked to large-scale tectonic processes, such as the India-Asia collision (e.g., Appel et al., 2012). The timing of secondary remanence acquisition can be estimated by comparing its direction to the predicted values from APWPs of Eurasia or India at the sampling localities.

3.2.3.2. State of the art. Most constraints on collision chronology based on remagnetization were obtained from the Tethys Himalayan Paleogene limestones exposed in the Tingri area. Besse et al. (1984) interpreted an age of ~50 Ma for the end of folding based on the intersection of paleolatitude defined by the secondary remanence and the Indian APWP. Appel et al. (1998) suggested that the secondary remanence isolated from the Zongpu limestones could have been acquired at any time between the Eocene and the Brunhes-Matuyama boundary, whereas Tong et al. (2008) inferred that it dated back to ~54 Ma and corresponded to the final suturing between India and Eurasia. Liebke et al. (2013) concluded that the latest possible time for secondary remanence acquisition was 48–37 Ma.

3.2.3.3. Conclusion and suggestions. Studies of remagnetization of Tethys Himalaya strata constrain the India-Asia collision only very poorly as older than ~54, or as older than 48–37 Ma. This approach can only provide a very rough minimum age for collision onset, limited by weak assumptions, including: 1) alteration of the primary remanence and acquisition of the secondary remanence were induced by thermal and/or chemical effects related to initial collision; 2) no post-collisional crustal shortening took place within Asia and the Himalaya; 3) no post-remagnetization tilting of strata occurred thereafter. In fact, mechanisms potentially causing remagnetization, including elevated temperature, tectonic stress, chemical alteration or secondary mineral growth, can act alone or in concert at much earlier or much later times than collision onset (Huang et al., 2015b,d,e). Large-scale upper crustal shortening took place both within Asia (600–750 km) and within the Himalaya (~900 km) since collision onset (Guillot et al., 2003; van Hinsbergen et al., 2011b; Robinson and Martin, 2014), and strong post-collisional deformation affected the Indus-Yarlung suture zone. Remagnetization can only provide very loose constraints on the initiation of collision, but the fact that it affects Tethys Himalayan limestones extensively must be taken into full account. Most paleomagnetic poles for the Tethys Himalaya were defined from limestones (e.g., Klootwijk and Bingham, 1980; Klootwijk et al., 1983; Appel et al., 1991; Patzelt et al., 1996; Yi et al., 2011), and it is urgent to check their reliability.

3.3. Other methods used to constrain collision onset

3.3.1. Faunal migration

3.3.1.1. Rationale. Before collision onset, animals and plants living in separate continents would show independent biological evolution. Once the distance between the opposite continental margins becomes small and they finally come into contact, faunas and floras can interact and start to experience the same evolution. The biological evolution recorded in the two continents thus provides a truly independent criterion to establish a minimum age for initial continental collision.

3.3.1.2. State of the art. Based on their study of terrestrial biota in the Deccan Plateau, Jaeger et al. (1989) suggested that India may have collided with Eurasia around the Cretaceous-Paleogene boundary. The distribution of terrestrial fossils and their migration history was reconstructed in far greater detail recently. Paleontology and molecular biology studies showed that several major orders of modern mammals, including adapisoriculid insectivores, whales, sea-cows, primates and horses originated in the Indian subcontinent (Bossuyt and Milinkovitch, 2001; Bajpai et al., 2008). Within Gondwana, Late Cretaceous *Eutheria* mammals were discovered only in India. Other animals, including some frogs, Caecilians and a few freshwater fishes such as arowanas and plants such as *Crypteroniaceae* and *Dipterocarpaceae* initially came from India. After a long period of isolation in the Cretaceous, during which the fauna and flora of peninsular India remained highly endemic and unique, exchange of biota with Asia began, and these species started to diffuse into Eurasia in the Cenozoic (the so called "Out-of-India" hypothesis; Bossuyt and Milinkovitch, 2001; Dutta et al., 2011).

The finding of mixed endemic (e.g., *cambaytheriid perissodactyls*) and northern families of continental mammals (*adapoid*, *omomyid* and *eosimiid primates*; *dichobunid artiodactyls* and *ailuravine rodents*; Bajpai, 2009) in the Vastan Lignite Mine of western India provided evidence that faunal exchange took place before ~54 Ma (Clementz et al., 2011). The Ghazij Formation in Pakistan also contains mixed India-Asia continental mammals, but younger than the Vastan biota (~50 Ma; Gingerich et al. 1997). The Subathu Formation of northern India and the Kuldana Formation of Pakistan dated at ~48 Ma yielded an even larger number of mixed species, including Asian-originated rhinoceroses, brontotheres and tapiroids (Thewissen et al., 2001).

Complementary evidence is provided by marine faunas and floras, which may record exchange at significantly earlier times than terrestrial faunas. Among larger benthic foraminifera, the southern Tethys Himalaya in the Late Cretaceous was characterized by *Orbitoides–Omphalocyclus* assemblages (Nicora et al., 1987; Willems et al., 1996), while *Lepidorbitoides*–dominated assemblages thrived along the southern Asian margin (Wan et al., 2002). Instead, upper Paleocene and lower Eocene strata along both margins yield similar assemblages including the same species of *Assilina*, *Daviesina*, *Discocyclina*, *Globanomalina*, *Lockhartia*, *Miscellanea*, *Nummulites* and *Ranikothalia* (Henderson et al., 2010; Zhang et al., 2013a; BouDagher-Fadel et al., 2015).

3.3.1.3. Conclusion and suggestions. The Vastan biota document that the first extensive migration of terrestrial faunas between India and Asia took place by the earliest Eocene (i.e., before ~54 Ma; Clementz et al., 2011). Mixing of shallow-marine benthic faunas occurred at somewhat earlier times before the end of the Paleocene (Garzanti et al., 1996). The earlier event documents the disappearance of the deep Neo-Tethys ocean separating the Indian and Asian margins, whereas the later event testifies to the demise of shallow seas replaced by a continuous terrestrial environment connecting the Asian upper plate with the Indian lower plate, as documented by the stratigraphic record (Garzanti et al., 1987; Najman et al., 2010; Hu et al., 2012). The faunal-migration approach is thus revealed as a reliable method able to provide an accurate minimum age for collision onset. Its application is however limited by the typically discontinuous fossil record in terrestrial settings, and by the difficult assessment of depositional age in the lack of marine index fossils.

3.3.2. Ultra-high pressure metamorphism

3.3.2.1. Rationale. High-pressure (HP) to ultra-high pressure (UHP) metamorphism of continental protoliths, documented by eclogitic parageneses locally containing minerals such as coesite or diamond that require subduction to mantle depths for their formation, is now widely documented in continental collision zones (Leech et al., 2005; Zheng, 2008; Guillot et al., 2008). The HP-UHP continental rocks are understood to have represented the leading edge of a hyper-extended margin, because unstretched thicker sections of buoyant continental crust would resist subduction to such great depths. The age of prograde HP-UHP metamorphism, together with the corresponding depth - inferred from metamorphic parageneses under the assumption of purely lithostatic pressure - and an estimate of subduction rate, can be used to calculate the age of initiation of continental subduction (Leech et al., 2005).

3.3.2.2. State of the art. In the western Himalaya, HP-UHP rocks occur just south of the Indus-Yarlung suture zone in the Kaghan area of Pakistan and in the Tso Morari area of India (Pognante and Spencer, 1991; Guillot et al., 1997; de Sigoyer et al., 2000; O'Brien et al., 2001). The pressure-temperature history experienced by these two UHP assemblages is similar: UHP peak pressures of up to 4.4–4.8 GPa at ~560–760 °C, followed by decompression and cooling to ~0.9–1.7 GPa at ~500–640 °C, and then reheating to ~0.8–1.2 GPa at ~650–720 °C (e.g., Wilke et al., 2015). Over the past decade two contrasting views have emerged: (1) UHP metamorphism occurred at 47–46 Ma at Kaghan, but earlier at 53–51 Ma at Tso Morari (e.g., Leech et al., 2005; St-Onge et al., 2013); (2) UHP metamorphism occurred at 47–46 Ma at both localities (e.g., Wilke et al., 2010; Donaldson et al., 2013). Under the assumption that the protoliths represented the very edge of the Indian margin, the first view would imply diachronous collision and a highly asymmetric collision geometry (with western India extending far to the north of Pakistan and colliding with Asia ~5 Ma earlier; Guillot et al., 2008) whereas the second view allows for quasi-synchronous collision along the western Himalaya.

UHP timing is best constrained in the Kaghan Valley by U-Pb dating of zircon rims containing coesite inclusions at 46.2 ± 0.7 Ma (2σ) (Kaneko et al., 2003), confirmed by further U-Pb zircon, Th-Pb allanite and $^{40}Ar/^{39}Ar$ phengite plateau ages (Wilke et al., 2010; Rehman et al., 2013). Tso Morari results have been more controversial because of larger uncertainties and/or smaller data sets. Leech et al. (2005) relied on 3 zircon U-Pb spot ages only, and similarly St-Onge et al. (2013) relied on a Concordia intercept age from zircon U-Pb spot dating that is largely determined by a single near-concordant age analysis only. Recently, a large dataset of U-Pb zircon ages indicated that peak eclogitic conditions were reached between ~47 and 43 Ma as in the Kaghan Valley, thus supporting a single UHP event and quasi-synchronous collision in the western Himalaya (Donaldson et al., 2013).

Similar HP-UHP rocks have not been reported so far in Tibet. Eclogites do occur close to the South Tibetan Detachment far south of the Indus-Yarlung suture in both Everest and Barwa regions (Groppo et al., 2007; Guillot et al., 2008; Zhang et al., 2013b), but their much younger metamorphic age (≤35 Ma) and prolonged exhumation suggests that metamorphism was unrelated to collision onset (Zhang et al., 2010).

3.3.2.3. Conclusion and suggestions. The ages of UHP metamorphism determined as described above were used to constrain the timing of India-Asia collision, under the assumption that the Indian margin at the time was subducting beneath Asia and not beneath an intervening intra-oceanic arc, as proposed for instance by Ali and Aitchison (2008) or Bouilhol et al. (2013). The reconstruction of P-T-t paths for HP-UHP rocks exposed in the western Himalaya coupled with an India-Asia convergence rate of 134 mm/a led Leech et al. (2005) to estimate that Indian continental crust first arrived at the Transhimalayan trench at 57 ± 1 Ma. After Warren et al. (2008) suggested that the exhumed UHP rocks were not derived from the edge of the continental margin but at least 200 km inboard of the ocean-continent transition, Leech et al. (2014) reassessed the onset of continental subduction at 61.1 ± 1.0 Ma. Based on their new dataset indicating an age for UHP metamorphism younger by ~7 Ma, Donaldson et al. (2013) favored instead onset of continental subduction at ~51 Ma or slightly later.

These calculations are based on a variety of assumptions and suffer from a variety of uncertainties. In particular the original palaeogeographic position of continental protoliths within the rifted margin cannot be reconstructed with any accuracy. With these limitations, HP-UHP metamorphism provides a valuable independent method to obtain minimum-age constraints on the onset of continental subduction.

3.3.3. Magmatism

3.3.3.1. Rationale. After transition from oceanic to continental subduction, magmatic activity on the upper plate may show different geochemical and isotopic signatures. Igneous rocks may even indicate changing magma sources, and provide indirect clues useful to broadly constrain the timing of initial continental subduction and thus of collision onset.

3.3.3.2. State of the art. Magmatic evidence was used in various ways to infer the timing of India-Asia collision. These include:

1) Presumed first appearance of collision-related magmatic rocks. In southern Tibet, the Linzizong volcanic rocks have been considered as the magmatic response to initial India-Asia collision (Mo et al., 2007), although numerous geochemical studies have shown that its lower part (Dianzhong Formation) maintains arc-related geochemical features. The beginning of Linzizong volcanism was dated as old as ~69 Ma (He et al., 2007), whereas recent SIMS U-Pb zircon ages documented that andesitic Dianzhong volcanism was active between 60.2 and 58.3 Ma in the Linzhou area (Zhu et al., 2015). This high-precision study led Zhu et al. (2015) to propose initial India-Asia collision at ~55 Ma, followed by slab break-off around 53 Ma and by pronounced magmatic flare up at 52–51 Ma.
2) End of "arc-like" magmatism. The youngest calc-alkaline volcanic rocks in the Gangdese batholith at ~40 Ma were proposed to constrain the timing of India-Asia collision (Aitchison et al., 2007a). Calc-alkaline magmatism, however, is recorded until 13 Ma at least (Chung et al., 2005; Ji et al., 2009), when Neo-Tethyan oceanic lithosphere had long disappeared.
3) Transition from I-type to S-type magmatism. The end of I-type Gangdese magmatism, supposedly related to oceanic subduction, and the appearance of S-type anatectic granites and migmatites in the Lhasa Block dated at the Eocene, were held to testify collision onset (Searle et al., 1987). St-Onge et al. (2010) observed that the Ladakh batholith documents granodioritic magmatism related to oceanic subduction until ~57 Ma, cross-cut by ~47 Ma leucogranites dykes, and consequently proposed collision onset between 57 Ma and 47 Ma. White et al. (2012) documented I-type hornblende-biotite granites as young as 31 Ma cross-cut by S-type leucogranites dated as 18 Ma, and thus favored initial collision between 31 and 18 Ma.
4) Abrupt changes of magma composition and presumed timing of hypothetic slab break-off events. Geochemical and geochronological studies of the Gangdese batholith have documented a magmatic flare-up at 52–51 Ma, with εHf and εNd values tending to become negative possibly due to contamination by continental crust (Ji et al., 2009; Chu et al., 2011). These observations were widely ascribed to break-off of Neo-Tethyan oceanic lithosphere (e.g., Lee et al., 2009), implying that initial collision may have occurred some 2–3 Ma before break-off and related magmatic flare-up (Zhu et al., 2015). Instead, Bouilhol et al. (2013) interpreted similar negative changes of εHf and εNd in the Kohistan-Ladakh batholith as the result of collision of India with the Kohistan-Ladakh arc at ~50 Ma, rather than with Asia. They argued that collision of India with Kohistan-Ladakh preceded the final India-Eurasia collision along the Shyok suture - separating Asia from Kohistan-Ladakh - at ca. 40 Ma. However, Borneman et al. (2015) recently documented that Asian-derived clastic wedges were deposited unconformably on Kohistan-Ladakh between ~92 and ~85 Ma, thus indicating a Late Cretaceous age for the Shyok suture. Upper Cretaceous syn-collisional clastic units from Ladakh to the northern Karakorum (Gaetani et al., 1993) indicate that the Kohistan-Ladakh arc was welded to Asia long before collision with India, suggesting that negative changes of εHf and εNd in the Kohistan-Ladakh batholith at ~50 Ma cannot be used to constrain the India-Asia collision timing.

3.3.3.3. Conclusion and suggestions. The geochemical and isotopic signatures of magmatism associated with oceanic and continental subduction vary according to different controls that can be seldom determined univocally. In the effort to constrain collision onset based on the magmatic record only, different researchers consequently made contrasting "ad hoc" hypotheses, often postponing collision age at times incompatible with other geological evidence. A more beneficial approach would be to invert terms, trying to understand the magmatic record in the light of more robust information provided by other methods.

3.3.4. Continental deformation

3.3.4.1. Rationale. Deformation associated with continental collision sweeps foreland-ward across the orogenic belt, starting with development of folds and thrusts along the suture zone (Carosi et al., 2016). During ongoing convergence, stress may also propagate toward the retro-side of the continental arc, were a retroarc fold-thrust belt and associated syn-collisional basin could form (Naylor and Sinclair, 2008). Theoretically, dating the earliest fold-thrust deformation, tectonic uplift, or metamorphic event close to the suture zone can provide a minimum-age constraint to collision onset.

3.3.4.2. State of the art. In the Zanskar Range, Tethys Himalayan strata were suggested to have been involved in south-verging fold-thrust deformation as early as the Late Cretaceous, interpreted as the time of obduction of the Spontang Ophiolite onto the Indian passive margin (Searle et al., 1997). Bonhomme and Garzanti (1991) obtained K-Ar illite ages of 47–44 Ma on epizonal lower Eocene red beds, interpreted to constrain at the middle Eocene the metamorphic event that affected the Tethys Himalayan succession beneath the ophiolite during the first stages of orogeny.

In northwestern Pakistan, thrusting of accretionary-wedge and trench strata onto the Indian passive margin between 66 Ma and 55.5 Ma indicated that Neo-Tethyan oceanic lithosphere disappeared there by the end of the Paleocene (Beck et al., 1995).

In southern Tibet, near the Indus-Yarlung suture zone, Ratschbacher et al. (1994) obtained K-Ar muscovite ages of 50.5–47 Ma for the Saga-Gyangze thrust in the Renbu area. Tectonic activity along this thrust was dated instead as 45–41 Ma near Saga by Ding et al. (2005), and as 70–60 Ma near Sangsang by Wang et al. (2016). The Yarlung-Zangbo mantle fault was inferred to have initiated around 63 Ma (Ding et al., 2005).

By studying leucogranites and their cross-cutting relationships with deformed Tethys Himalayan strata, Aikman et al. (2008) suggested that crustal shortening began before 44 Ma. Smit et al. (2014) obtained Lu-Hf ages of 54–52 Ma and 51–49 Ma on garnets grown in the Mabja and Kangma metamorphic complexes of the northern Tethys Himalaya, which were related to compressive metamorphic deformation at middle crustal levels during ongoing India-Asia collision.

In the southern Lhasa Block, the base of Linzizong volcanic rocks dated as ~60 Ma in Liuzhou and as ~68 Ma in Maxiang near Lhasa (Zhu et al., 2015) truncates with angular unconformity the underlying Permian to lower Upper Cretaceous strata. Such strong deformation, interpreted as the result of India-Asia collision by Mo et al. (2003, 2007), took place between ~90 Ma and ~60 Ma. However, the pre-Linzizong succession displays large open folds with subvertical cleavage comparable to structures observed in the Andean range and thus probably acquired during the oceanic-subduction stage that preceded the India-Asia collision (Burg et al., 1983; Kapp et al., 2007).

3.3.4.3. Conclusion and suggestions. Most available evidence suggests that tectonic shortening and crustal thickening in the Himalayas began between 54 and 44 Ma. This indicates that orogeny was well underway at that time, which represents a minimum-age constraint for collision onset. Continent-continent collision, however, is not the only possible cause for compressional tectonic deformation. Alternative geological processes were invoked, including early events of ophiolite obduction (Searle et al., 1997) or arc-continent collision (Aitchison et al., 2007b), which will be discussed in detail in subsequent sections of this article.

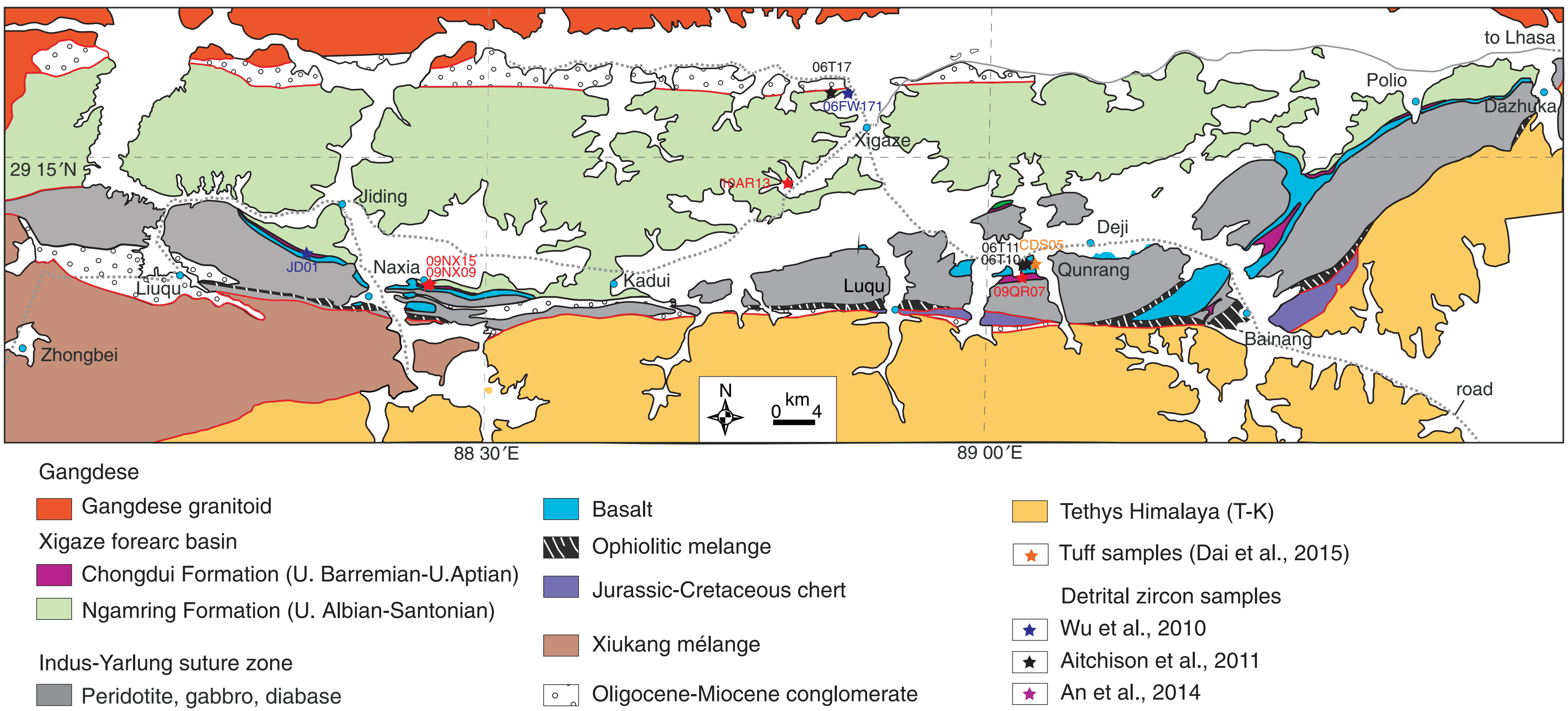

Fig. 8. Simplified geological map of the Xigaze area (modified from Wang et al., 1984; Dai et al., 2015) showing the locations of tuff and sandstone samples analyzed for U-Pb zircon geochronology as displayed in Fig. 11.

4. The Paleogene arc-continent collision hypothesis

According to the island arc–continent collision hypothesis, proposed by Aitchison et al. (2000, 2007b), India collided first with an intra-oceanic island arc at ~55 Ma, and was finally welded with Asia only at ~34 Ma. To evaluate this hypothesis, which has been rebutted by Garzanti (2008), both supporting and contradicting paleomagnetic, igneous, sedimentary and palaeontological evidence should be examined carefully. In their vision of the geology of the Indus-Yarlung suture zone, Aitchison et al. (2000), Aitchison and Davis (2004), Aitchison et al. (2007b) identified three distinct "terranes": Zedong, Dazhuqu and Bainang (Fig. 8). The "Zedong terrane", exposed for mere ~25 km^2 in the Zedong area, was considered as a possible intra-oceanic island arc similar to the Kohistan arc in the Pakistan Himalaya. The "Dazhuqu terrane" includes the Yarlung-Zangbo Ophiolite, the ophiolitic mélange and the Chongdui Formation in the Xigaze area, whereas the "Bainang terrane" consists of Jurassic to Lower Cretaceous red bedded chert exposed south of the ophiolites (Ziabrev et al., 2004; Xialu Chert Formation by Matsuoka et al., 2002).

4.1. The Zedong terrane: remnant of an intra-oceanic arc?

The "Zedong terrane", a thrust slice within the Great Counter Thrust system, lies in fault contact with an ultramafic ophiolite massif in the south and with the upper Oligocene–lower Miocene Luobusha Conglomerate in the north, where it is partially covered by alluvial deposits of the Yarlung River. This suite of volcanic and volcaniclastic rocks comprising basaltic-andesites, andesites, andesitic breccias with rare dacites and other intrusives (hornblendite, gabbro, granite; Aitchison et al., 2000) is dated gechronologically at the Late Jurassic (160–155 Ma; McDermid et al., 2002; Zhang et al., 2014).

4.1.1. Evidence in favor

Initially ascribed to the tholeiitic series, some Zedong lavas were subsequently classified as shoshonites because of their high K_2O content (Aitchison et al., 2000, 2007a). And because shoshonites occur commonly in modern intra-oceanic arcs, the Zedong terrane was considered as the remnant of an intra-oceanic arc originally located within the Neo-Tethys Ocean (Aitchison et al., 2007a).

4.1.2. Evidence against

Several lines of evidence clash with the interpretation of the "Zedong terrane" as the remnant of an intra-oceanic arc.

(1) Recent studies revealed that Zedong volcanic rocks are extensively altered, and that their content in mobile elements such as K and Na cannot be used for rock classification. The distribution of immobile elements (e.g., Th–Co and Zr/TiO_2–Nb/Y) points to the calc-alkaline rather than to the shoshonite series (Zhang et al., 2014).
(2) No ophiolite is found between the "Zedong terrane" and the Gangdese arc; ophiolites crop out only to the south of the "terrane".
(3) Geochemical signatures of Zedong volcanic rocks are similar to those of Jurassic volcanic rocks exposed along the southern margin of the Lhasa Block, widely interpreted as part of the Asian active continental margin (Zhu et al., 2008).
(4) In Aitchison's model intra-oceanic subduction was continuous from the Jurassic to the early Paleocene (Aitchison et al., 2000, 2007b), whereas Zedong volcanic rocks were erupted in a very short period of time (~5 Myr only; McDermid et al., 2002; Zhang et al., 2014). Presuming that all younger volcanic rocks have disappeared entirely is hardly viable.
(5) Zedong volcanic rocks are restricted to a very limited area, and were never reported from other localities along the Indus-Yarlung suture zone in Tibet.

4.1.3. Conclusion

We conclude that evidence against overwhelms evidence in favor, and consequently the "Zedong terrane" may not represent a remnant intra-oceanic arc. Rather, it may represent a slice of the active continental-margin arc developed at the southern margin of the Lhasa Block (Zhang et al., 2014).

4.2. The Dazhuqu terrane: part of an intra-oceanic subduction system?

Dispute has long existed on the origin of the Yarlung-Zangbo Ophiolite (e.g. Hébert et al., 2012; Wu et al., 2014b; Zhang, 2015). In this section we shall focus specifically on the claim that the ophiolites together with the overlying Chongdui chert and sandstones may represent an intra-oceanic "Dazhuqu terrane" (Aitchison et al., 2000).

4.2.1. At which paleolatitude did the Yarlung Zangbo Ophiolite form?

Paleomagnetic studies have provided contrasting results for the Yarlung-Zangbo Ophiolite. Pozzi et al. (1984) calculated a paleolatitude of 10–20°N for the chert member of the Chongdui Formation and limestones of the Sangzugang Formation, concluding that the Yarlung Zangbo Ophiolite formed near the Lhasa Block. In contrast, Abrajevitch et al. (2005) estimated a near-equatorial paleolatitude for Chongdui radiolarites, siliceous mudstones and volcaniclastic rocks capping the ophiolite, a result used widely to support an intra-oceanic setting (e.g., Aitchison et al., 2007b; Hébert et al., 2012). Both datasets were however small, and potential artefacts associated with inclination shallowing and remagnetization were not assessed critically. Recently, Huang et al. (2015c) re-investigated cherts and turbiditic sandstones of the Chongdui Formation at Chongdui, Bainang and Sangsang, obtaining paleolatitudes of 11.2 ± 3.5°N, 7.9 ± 5.8°N, and 8.6 ± 3.5°N, respectively. Positive fold tests indicated pre-folding and probably primary origin of the isolated ChRMs. The flat distribution of the virtual geomagnetic poles (VGPs) corresponding to the ChRMs from all three sections, however, indicates that ChRMs were biased by inclination shallowing. Although the Chongdui and Bainang datasets are too small to apply a shallowing correction, two independent correction methods applied to Sangsang data yielded consistent paleolatitudes of 16.2°N [13°N, 20.9°N] and 16.8°N [11.1°N, 23.3°N], which are statistically indistinguishable from the Early Cretaceous (~120 Ma) paleolatitudes of the Lhasa Block (see Section 3.2.1.2; Sun et al., 2010; Chen et al., 2012; Ma et al., 2014; Yang et al., 2015b; Li et al., 2016). Such paleomagnetic evidence supports formation of the Yarlung-Zangpo Ophiolite adjacent to the Lhasa Block, in the Gangdese forearc rather than near the equator ~2500 km south of Eurasia (Huang et al., 2015c; Xiong et al., 2016).

4.2.2. Stratigraphy of the Chongdui Formation

The lower part of the Chongdui Formation consists of grey chert with varying thicknesses deposited stratigraphically on pillow basalts of the Yarlung-Zangbo Ophiolite, as well exposed in the Qunrang, west Dazhuka, Pilo and Naxia sections and reported since Sino-French Expeditions in the 1980s (Fig. 5H; Nicolas et al., 1981; Girardeau et al., 1984, 1985a,b; Wu, 1984; Dai et al., 2013; An et al., 2014). The contact between the lower chert member and the overlying sandstone member, clearly exposed in the Qunrang, Naxia and Polia sections, is also stratigraphic (Nicolas et al. 1981; Girardeau et al. 1984, 1985a,b; Wu, 1984; Wang et al., 1999, 2012; An et al., 2014) (Fig. 5G; Fig. 9).

The chert member of the Chongdui Formation contains abundant mid-Cretaceous radiolaria that - based on correlation with the radiolarian zonation of the Great Valley sequence in California - were dated at the late Albian-early Cenomanian (Marcoux et al., 1982; Wu, 1984; Li and Wu, 1985). Correlation with the mid-Cretaceous radiolarian range zones established in pelagic limestone in Italy (O'Dogherty, 1994) allowed Ziabrev et al. (2003) to reassign the member to a relatively short time interval between the late Barremian and the late Aptian (127–113 Ma). This age is consistent with the published radiometric ages of underlying magmatic rocks (124–126 Ma ^{39}Ar-^{40}Ar hornblende

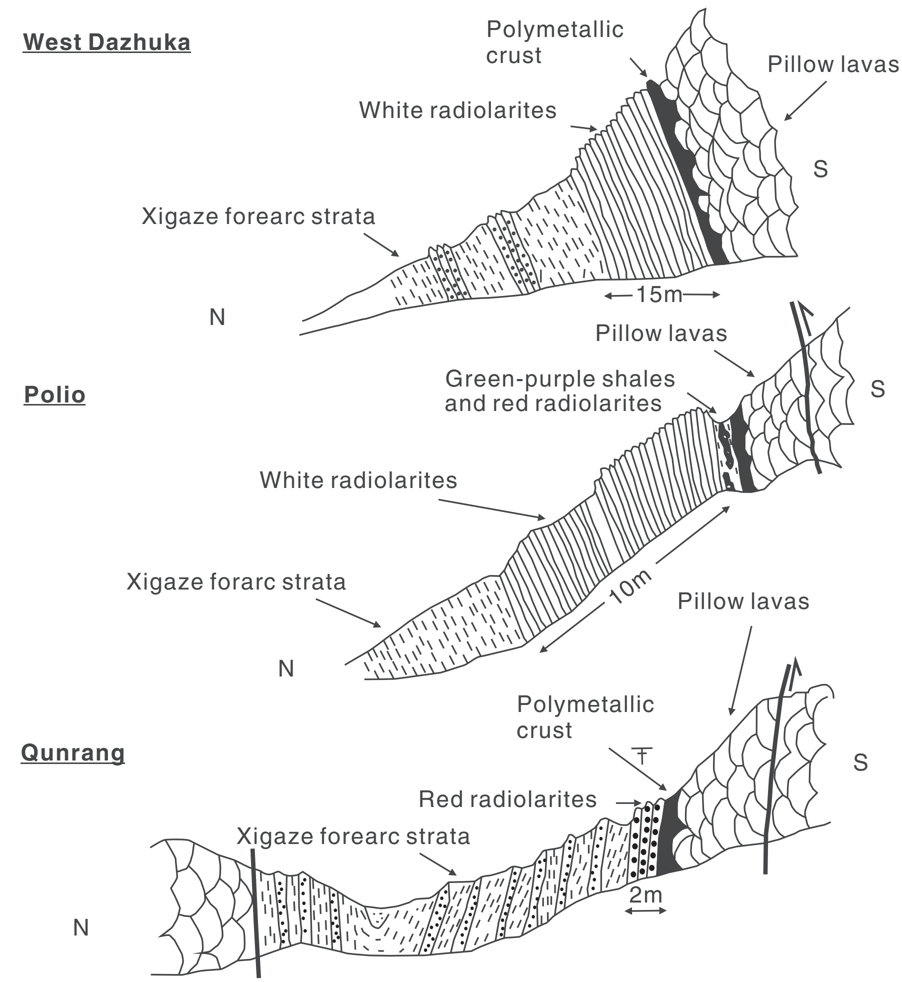

Fig. 9. Stratigraphic contact between the Chongdui Formation and underlying pillow lavas atop the Indus-Yarlung Zangbo ophiolitic sequence as observed in the Qunrang, Polio, Dazhuka, and Naxia sections.
(After Girardeau et al., 1984, p.159).

age from Qunrang and Bainang: Guilmette et al., 2009; 126 Ma U-Pb zircon age from gabbro in Qunrang and diorite in Dazhuka: Malpas et al., 2003). A tuff from the overlying Chongdui sandstone member was dated as 116–114 Ma (Dai et al., 2015; Huang et al., 2015c), constraining the chert member as older than the Albian.

4.2.3. Relation between the Chongdui and Ngamring formations

The Chongdui Formation was first distinguished from the Ngamring Formation because (Cao, 1981): (1) the Chongdui Formation includes a chert member deposited directly on basalt, whereas the Ngamring Formation consists of submarine fan deposits; (2) the Chongdui Formation was deposited below the calcite compensation depth, whereas the Ngamring Formation contains abundant interbedded marls and was thus deposited at shallower depths; (3) the Chongdui sandstone member is dominated by lithic fragments (>80% QFL), including chert (52%), serpentinite (30%) and basalt (18%), and contains pyroxene, olivine and magnetite (sandstones were thus called improperly as “ophiolitic greywackes”; Wu, 1984; Yin et al., 1988).

Different tectonic settings have been proposed for the Chongdui sandstone member (Fig. 10). Turbiditic deposition in abyssal environments and ophioliticlastic composition suggested a trench or trench-slope-basin setting and subduction-complex provenance (Cao, 1981; Yin et al., 1988). Other authors suggested that the Chongdui and Ngamring turbidites were both deposited in the Xigaze forearc basin but in different depocenters (Wu, 1984; Wang et al., 1999, 2012). Different compositions suggested that the Chongdui sandstone member

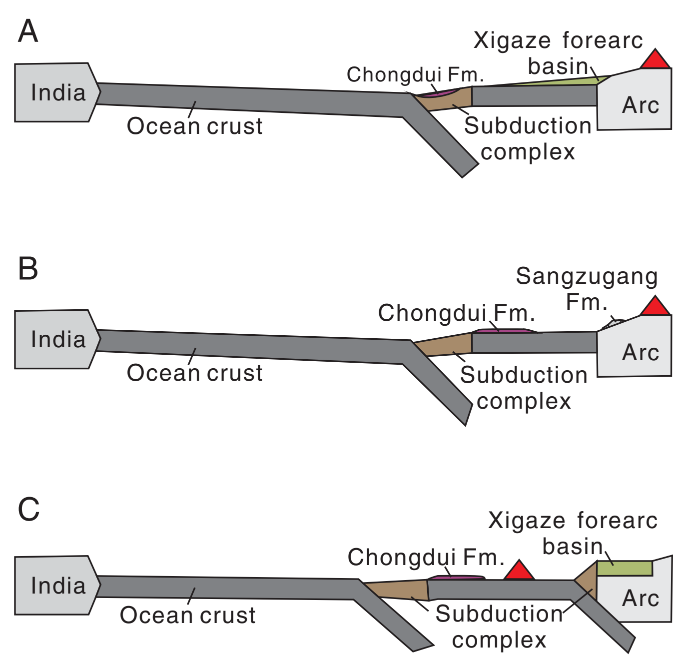

Fig. 10. The three alternative depositional settings proposed for the Chongdui Formation discussed in the text: A) trench or trench-top basin (Cao, 1981; Yin, 1988); B) base of the Xigaze forearc basin (Wu, 1984; Wang et al., 1999, 2012) C) forearc basin of an intraoceanic arc (Aitchison et al., 2000).

might have been deposited in a trench-slope basin and derived from ophiolites obducting in the south during the initial stages of Neo-Tethyan subduction, whereas the Ngamring Formation was deposited in the Xigaze forearc basin and fed from the Gangdese arc in the north (Wu, 1984). These models required contemporaneous obduction, uplift and erosion of an ophiolite complex.

Aitchison et al. (2000) suggested that the Chongdui sandstone member is genetically unrelated to the Ngamring Formation, based on the following evidence: (1) the Chongdui sandstone member contains volcaniclastic turbidites and abundant devitrified felsic tuffs, with significant quantities of detrital magnetite locally; (2) the possibility of a depositional contact between the two units would be precluded if the top of the Chongdui sandstone member was Albian in age, younger than the overlying Ngamring turbidites; (3) a volcanogenic component appeared in the Xigaze forearc basin only around the Aptian-Albian boundary, thus later than deposition of the main body of the Chongdui sandstone member. On this basis they concluded that also the Chongdui sandstone member may have been deposited in a forearc basin, but distinct from the Xigaze forearc basin (Aitchison et al., 2000 p.236). In Fig. 2B of Aitchison et al. (2000), however, the Chongdui sandstone member exposed in the Jiding-Naxia area (Sagui) is drawn as part of the Xigaze forearc basin, in fault contact with the ophiolite.

This apparently complicated issue has been recently clarified by rich U-Pb detrital-zircon age and Hf isotope datasets from Chongdui and Ngamring turbidites in various locations, which have provided firmer constraints on the provenance of both units (Fig. 11; Wu et al., 2010; Aitchison et al., 2011; An et al., 2014; Dai et al., 2015). In the Jiding and Naxia sections, the Chongdui sandstone member overlies the lower chert member conformably (Fig. 5G) and displays age peaks at 115 and 106 Ma, and positive $\varepsilon_{Hf}(t)$ values from +5 to +20% (left column in Fig. 11). The same signatures are shown by lower Ngamring sandstones near Xigaze. Three sandstone samples from the Qunrang type-section of the Chongdui sandstone member near Bainang display age peaks at 125 Ma, 115 Ma, and 110 Ma, respectively, and positive $\varepsilon_{Hf}(t)$ values similar to those of the Ngamring Formation (Fig. 11; Table DR4). In contrast, a tuff from the base of the member dated at 116–114 Ma yielded two groups of $\varepsilon_{Hf}(t)$ values, one between −6.5 and +1.4 and the other between +12.1 and +22.7 (Dai et al., 2015). Our own tuff sample (09QR07, Table DR4) from the base of the Chongdui Formation shows high positive $\varepsilon_{Hf}(t)$ values ranging from +11.8 to +18.6, except one grain yielding a value of +1.6 (Fig. 11).

4.2.4. Conclusion

The Chongdui chert member lies in primary depositional contact with the underlying pillow basalts and makes a stratigraphic transition to the overlying turbidites, as already documented by Sino-French Expeditions in the 1980s and confirmed by recent studies of the Qunrang, Naxia and Polio sections. The ages of the underlying ophiolite, of the chert and sandstone members of the Chongdui Formation, and of the overlying Ngamring turbidites form an orderly sequence fully compatible with a normal stratigraphic succession. Chongdui turbidites are similar to Ngamring turbidites as facies, composition and provenance. Both were deposited in submarine fans fed by the Lhasa Block. Integrated stratigraphic, paleomagnetic and provenance studies indicate that the Xigaze Ophiolite represents the stratigraphic basement of the Xigaze

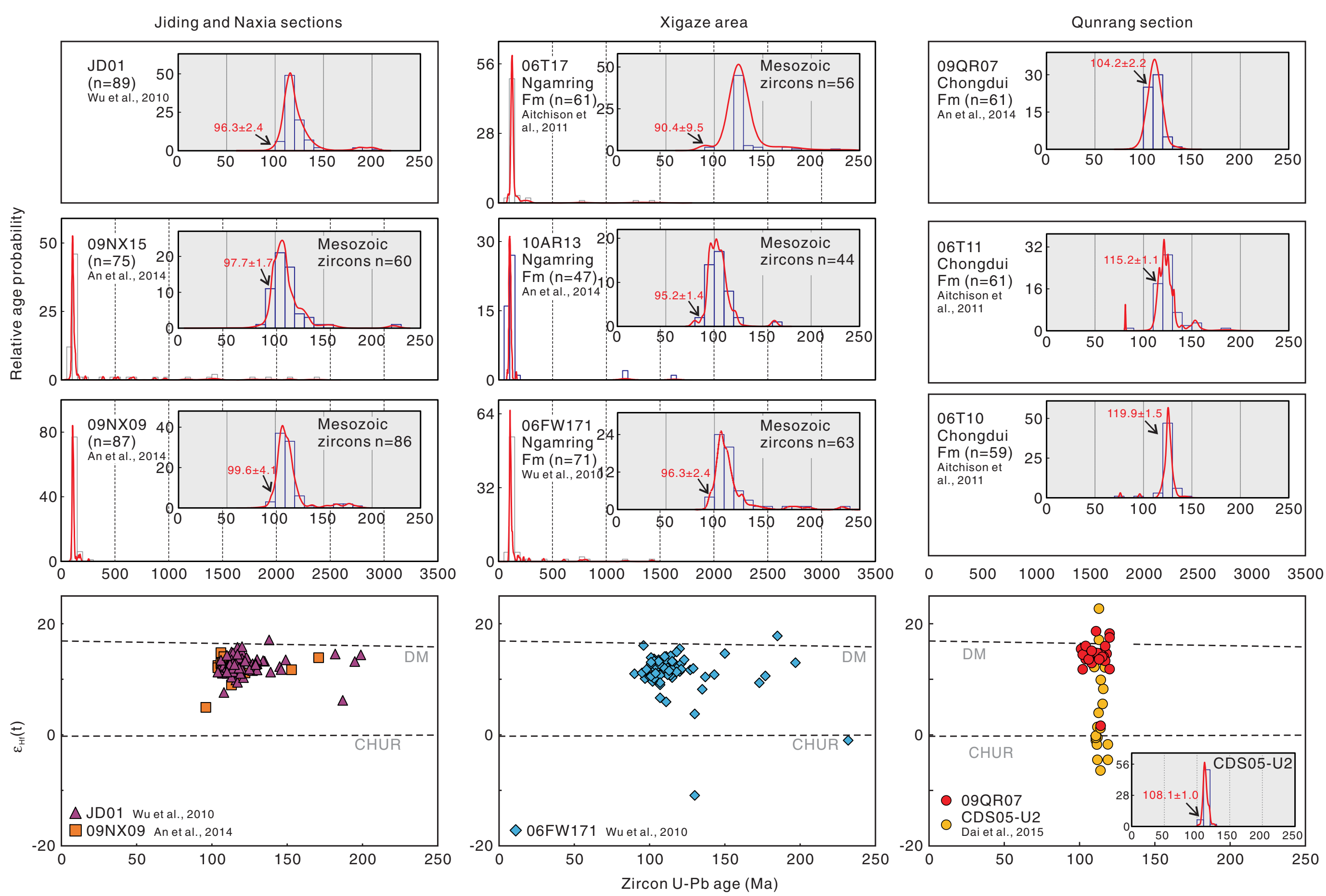

Fig. 11. Age spectra and Hf isotopic signatures of detrital zircons from the Chongdui sandstone member, compared to those of the lower member of the Ngamring Formation of the Xigaze forearc basin. Left: Jiding-Naxia area; Middle: Xigaze city-Qumei area; right: Qunrang section (locations shown in Fig. 8). Data sources: Wu et al. (2010), An et al. (2014), Aitchison et al. (2011), Dai et al. (2015), and this study.

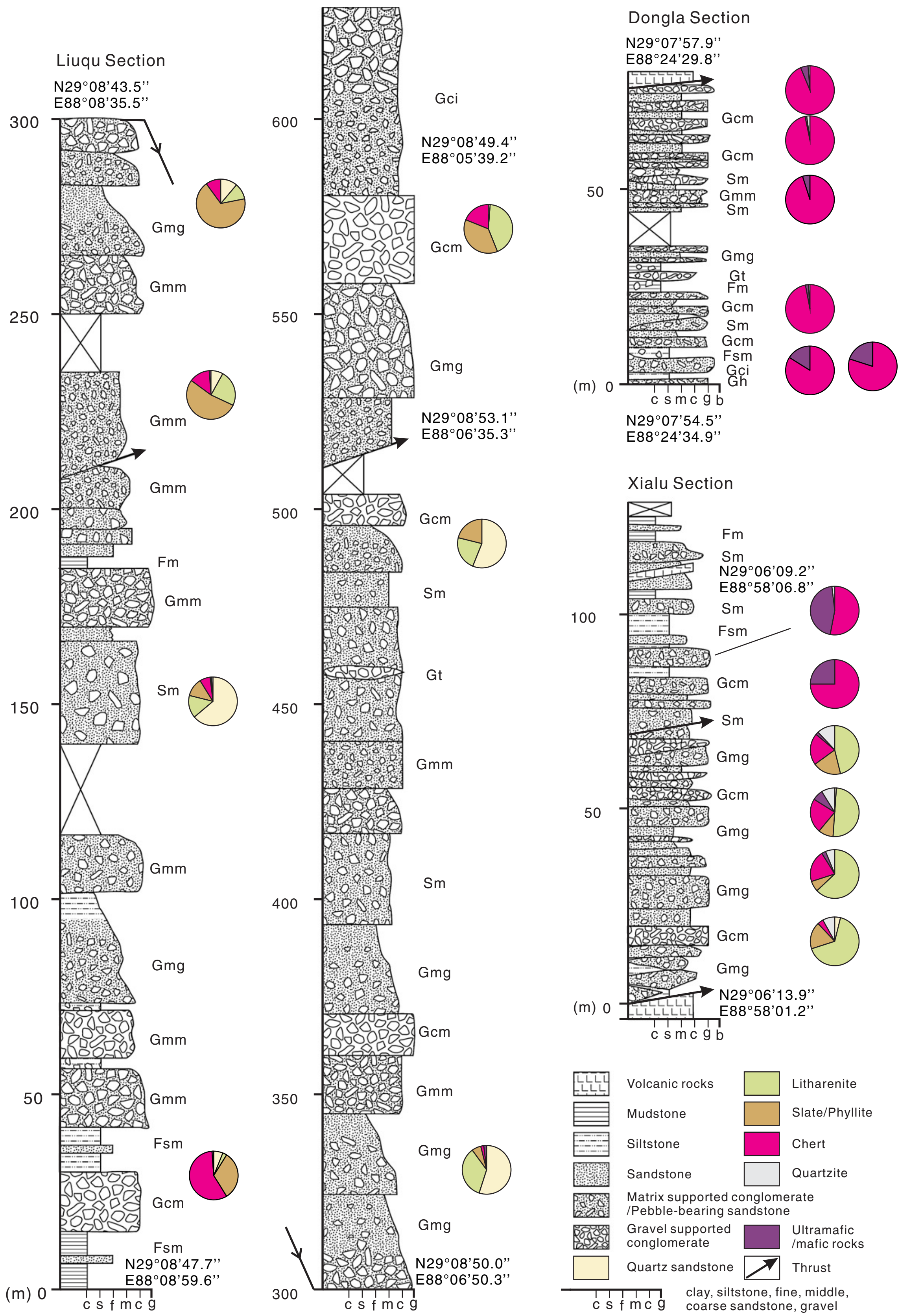

Fig. 12. Measured stratigraphic sections of the Liuqu Conglomerate, showing lithofacies and gravel composition. (A) Liuqu section; (B) Xialu section; (C) Tongla section. GPS positions are provided. Lithofacies are described in full in Table DR2.

forearc basin, and that the Chongdui Formation documents the transition from abyssal sedimentation during the earliest starved stage of forearc-basin sedimentation to the initial filling stage continued with thicker turbidites of the Ngamring Formation (An et al., 2014). Consequently, the Chongdui chert and sandstone members, together with the underlying ophiolites ("Dazhuqu terrane" of Aitchison et al., 2000) did not represent a separate intra-oceanic domain.

4.3. The Liuqu Conglomerate: record of arc-continental collision?

The up to 3.5 km-thick Liuqu Conglomerate is exposed discontinuously over a distance of 150 km along the Indus-Yarlung suture zone in the Xigaze area. Traditionally, the Liuqu Conglomerate was considered as a Paleogene "molasse" deposited after the India-Asia collision and documenting uplift and erosion of the Indus-Yarlung suture zone (Yin et al., 1988). In the models of Aitchison et al. (2000) and Davis et al. (2002), however, it was interpreted as the product of collision between India and an intra-oceanic arc.

4.3.1. Facies and composition

Poorly sorted, matrix- to clast-supported conglomerates, pebbly sandstones and mudrocks are predominantly reddish or locally light green. Angular to well rounded clasts are typically 2–20 cm in diameter with rare boulders up to 1 m or even larger, indicating deposition in proximal alluvial-fan to braided-river environments (Davis et al., 2002; Wang et al., 2010a; Leary et al., 2016b). Clasts include basalt, gabbro, serpentinite, radiolarian chert of various color, as well as quartzarenite, litharenite, phyllite and slate, indicating provenance from both the ophiolitic suture and units of Indian affinity. Gravel derived from the Gangdese arc massif, such as granite or intermediate-felsic volcanic rocks, was not detected (Davis et al., 2002; Wang et al., 2010a).

4.3.2. Depositional age

The age of the Liuqu Conglomerate is poorly constrained because age-diagnostic fossils are lacking. Rich assemblages of plant fossils found in the Liuqu type-locality were ascribed to the middle or late Eocene (Tao, 1988; Fang et al., 2004), whereas palynological studies suggested a late Paleogene age (Wei et al., 2009). The youngest reported U-Pb detrital-zircon ages are ~58 Ma (Wang et al., 2010a) and 32 Ma (Leary et al., 2016b), representing maximum depositional ages. The numerous north-directed thrust faults disrupting the succession, associated with the Great Counter Thrust or Renbu–Zedong thrust system (Yin et al., 1994, 1999), are cut in turn by felsic dikes dated at 18.3 ± 2.7 Ma (Yin et al., 1994; Williams et al., 2001), thus representing a minimum depositional age (Davis et al., 2002). More recently, apatite fission-track and apatite and zircon (U–Th)/He data have constrained deposition of the Liuqu Conglomerate to the latest Oligocene–Early Miocene (Li et al., 2015a). An early Miocene age (19–20 Ma) was favored by Leary et al. (2016b) based on biotite $^{40}Ar/^{39}Ar$ data, detrital zircon fission-track analyses, and $\delta^{13}C$ values of soil carbonates.

4.3.3. Is Asian-derived detritus indeed absent?

Lack of Gangdese-derived gravels have led to interpret the Liuqu Conglomerate as a product of collision between India and an intra-oceanic arc, predating the final collision with Asia (Aitchison et al., 2000; Davis et al., 2002). The Liuqu Conglomerate, however, contains many pebbles of lithic-rich sandstones (e.g., in the lower part of the Xialu section or in the upper part of the Liuqu section; Fig. 12; Table DR3), which yielded in turn detrital zircons with the typical Cretaceous U-Pb age population with positive εHf(t) values of Gangdese affinity (Fig. 13; Table DR4). Interbedded sandstone layers also yielded detrital zircons with Cretaceous–early Eocene U-Pb ages and positive εHf(t) values (Wang et al., 2010a; Leary et al., 2016b) (Fig. 13). These grains were interpreted to be recycled from Xigaze forearc-basin strata (Li et al., 2015a), or from the Xiukang accretionary mélanges along the southern Asian margin prior to India-Asia collision (Leary et al., 2016b).

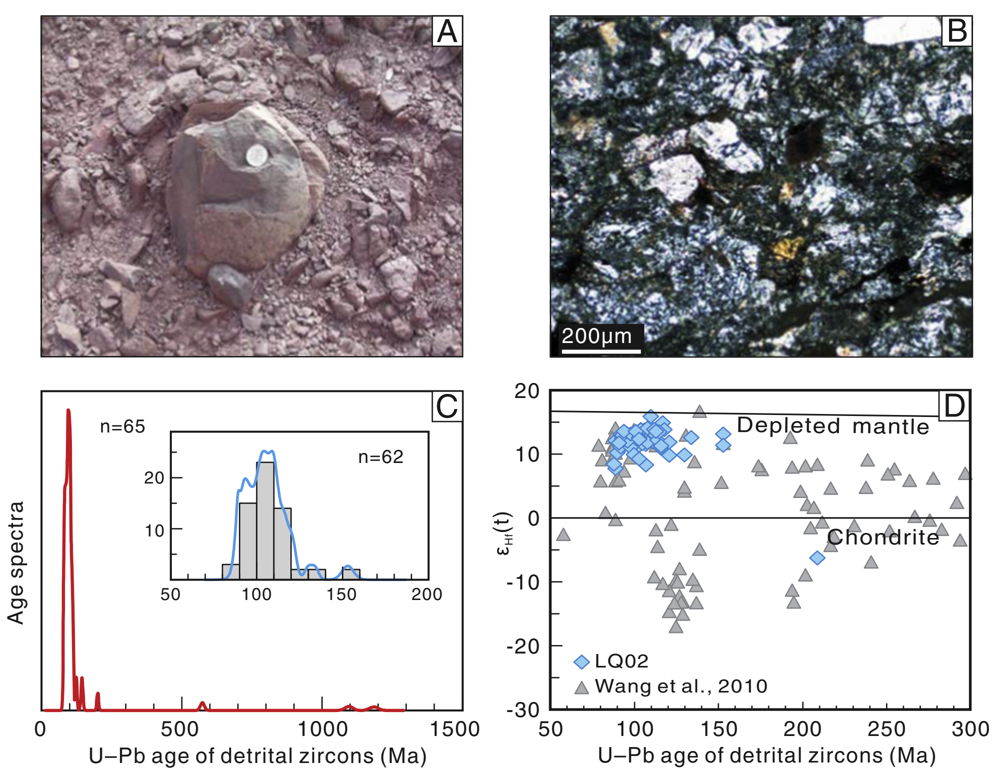

Fig. 13. Field (A) and thin-section photographs (B) of a cobble of feldspatho-lithic volcaniclastic sandstone (LQ02) in the Liuqu Conglomerate. (C) U-Pb age population of detrital zircons. (D) Plots of εHf(t) vs. U-Pb age of detrital zircons within the volcaniclastic sandstone cobble.

4.3.4. Conclusion

Detrital-zircon geochronology and geochemistry indicates that, whether primarily derived from the Gangdese arc or recycled from the Xigaze forearc basin and/or the Xiukang accretionary mélange, detritus from Asia is present in the Liuqu Conglomerate. Provenance data thus concur with structural and thermochronological evidence to indicate that the Liuqu Conglomerate was deposited long after initial India-Asia collision, thus ruling out the Paleogene India-intraoceanic arc collision model.

4.4. The Paleogene arc-continent collision hypothesis rejected

Aitchison et al. (2000, 2007b) proposed their Paleogene (~55Ma) intraoceanic arc-continent collision model based on: (1) paleomagnetic evidence; (2) end of marine sedimentation; (3) beginning of continental "molassic" sedimentation; (4) regional sedimentation patterns and major unconformity with absence of upper Eocene-Oligocene strata in the Lesser Himalaya; (5) end of calc-alkaline magmatism along the southern Asian margin; (6) initiation of major collision-related thrust systems.

Points 1, 2, 3, 5 and 6 have been discussed in Section 4 above, where we showed how these approaches can provide only minimum-age and/or unrobust criteria to constrain collision onset. For instance, Aitchison et al. (2000, 2007b) followed Searle et al. (1987), stating that "*the timing of terminal collision of the two plates is deduced from the ending of marine sedimentation in the Indus Suture Zone* (*ISZ*)". This criterion is invalid, because the final disappearance of marine seaways may and does postdate the first continent-continent contact by several Myr to tens of Myr, in the Himalayas as in other collision orogens, depending on diverse factors including basin width, rate of basin filling or shortening rate (e.g., Sinclair, 1997; Allen and Allen, 2005; Garzanti and Malusà, 2008).

As far as point 4 is concerned, the origin of the late Eocene-Oligocene unconformity widely documented in foreland-basin strata (e.g., in the Tansen Group of the Nepalese Lesser Himalaya, ~225 km south of the Indus-Yarlung suture zone) is indeed poorly understood. Five hypotheses have been proposed (fig. 25 and p. 98–99 in Yin, 2006): (1) passage of a flexural bulge ; (2) uplift of northern India above the hot thermal anomaly induced by mantle upwelling caused in turn by break-off of the Neo-Tethyan slab in the early stage of India-Asia collision; (3) enhanced forebulge caused by steeper geometry of the Neo-Tethyan slab still attached to India until after the early Miocene; (4) contraction and broad uplift of the Indian foreland basin in the Oligocene at a larger scale than occurred subsequently to the Shillong Plateau; (5) climatic change (Clift and VanLaningham, 2010). The late Eocene-Oligocene unconformity may thus have different causes, and can not be taken as supporting evidence to the Paleogene introcenaicarc-continental collision.

The available stratigraphic, provenance, paleomagnetic and tectonic evidence, as summarized above, contradicts the existence of independent terranes such as the "Zedong terrane" or "Dazhuqu terrane" within the Indus-Yarlung suture zone, and does not support the Paleogene intraoceanic arc-continent collision model. No more complex scenario than the one involving the passive continental margin of India and the active continental margin of Asia is needed to explain the geological evolution of the nascent Himalayas.

5. The Greater India Basin hypothesis

Plate reconstructions based on magnetic anomalies suggest that the convergence between India and Asia since 52 Ma amounts to ~3600 km, whereas the estimated upper crustal shortening amounts to only ~2350 km. To explain this "shortening deficit", van Hinsbergen et al. (2012) proposed that the Himalaya was separated from cratonic India by a "Greater India Basin" (GIB), formed as a result of N-S extension between 120 and 70 Ma. Following early Paleogene collision between the Tethys Himalaya micro-continent with Asia, such a 2675 ± 700 km-wide basin would have been subducted beneath the Himalaya from 50 to 25 Ma, and the "real" hard collision between India and Asia + Tethys Himalaya would have occurred only as late as 25–20 Ma.

5.1. Geological evidence, in favor or against?

The GIB hypothesis was introduced ad hoc to explain the discrepancy between total convergence and crustal shortening, which in the lack of supporting geological evidence makes it specious. Early Cretaceous extension associated with widespread volcanism between 140 and 110 indeed occurred along the northern Indian margin (e.g. Garzanti, 1999; Hu et al., 2010, 2015b). No evidence however indicates that an oceanic basin such as the GIB was generated at that time. No remnant of related ophiolite and/or arc-trench system are found within or below the Greater Himalaya (Aitchison and Ali, 2012), where the GIB suture should be located. No trace is left anywhere of the syn-rift and post-rift successions that should have been accumulated throughout the Cretaceous and Paleogene on the supposedly newly formed continental margins along both sides of the GIB.

Furthermore, the Patala and Balakot formations of northern Pakistan, the Subathu Formation of India and the Bhainskati Formation of Nepal in the Lesser Himalaya were sourced in various proportions from the Indus-Yarlung suture zone and the Tethys Himalaya in the early and middle Eocene (Critelli and Garzanti, 1994; DeCelles et al., 1998, 2004; Najman and Garzanti, 2000; Najman et al., 2005; Jain et al., 2009; Ravikant et al., 2011). This proves that in the early-middle Eocene the Tethys Himalaya was joined to the Lesser Himalaya (DeCelles et al., 2014), which rules out the existence of a Greater India Basin at that time. Nd and Sr isotopic evidence that a paleo-Indus fluvio-deltaic system was transporting Asian-derived detritus southward all across northern India by 50 Ma represents additional proof against the GIB hypothesis (Zhuang et al., 2015). Lack of separation between the Tethys Himalaya and India is indicated independently by the high-Ti (>10 wt% TiO_2) Cr-spinels found in the uppermost Cretaceous and lower Paleocene Zhepure Shanpo and Jidula formations, interpreted as partly sourced from the Deccan large igneous province (Garzanti and Hu, 2015).

5.2. Where should we look for a "cryptic" suture?

A testable prediction of the GIB hypothesis (van Hinsbergen et al., 2012) is that the Himalayan mountains should host a second 'GIB' suture to the south of the Indus-Yarlung suture zone, predicted to occur within or at the base of the Greater Himalaya. Alternatively, the GIB suture would be located within the Tethys Himalaya (DeCelles et al., 2014). In both settings, no ophiolite and/or arc-trench-system record has ever been observed. The GIB suture should thus be "cryptic" and entirely eroded away.

A cryptic or eroded suture would be most difficult to identify within the Greater Himalaya and particularly along its bounding faults, as the Greater Himalaya represents a stack of high-grade rocks separated from lower-grade sequences above and below along major shear zones (Yin, 2006). Greater Himalayan rocks were dominantly constructed by forward-propagating thrust stacking of material scraped off of the down-going Indian plate during the late Eocene to Miocene (e.g., Le Fort, 1975; Corrie and Kohn, 2011; He et al., 2015; Carosi et al., 2016). Such a coherent structural evolution (Bollinger et al., 2004; He et al., 2015; Larson et al., 2015; Yu et al., 2015) has no obvious geometric element within which to 'hide' a suture.

Even more unlikely is a "cryptic" suture located within the Tethys Himalaya, a fully developed stratigraphic succession where nothing comparable to such an imaginary structure has ever been reported (Sciunnach and Garzanti, 2012). The western Himalaya represents a particularly strong challenge to the GIB hypothesis because the Greater Himalaya is not exposed across much of the region. In particular, across the Chamba Himalaya (~77°E) and from Srinagar (~75°E) to the west, locally excepting the syntaxis, Tethys Himalayan strata can be traced continuously from the Indus suture to the Lesser Himalaya (DiPietro and Pogue, 2004; Yin, 2006; Webb et al., 2011). To the west of Srinagar, Tethys Himalaya strata are deposited directly atop Lesser Himalayan rocks, and both sequences are semi-continuous (i.e., can be traced across restorable faults) from the Indus suture to the undeformed Indian craton. The stratigraphic and structural records that demonstrate this continuity are rich and well-established. Most key lithologies are Phanerozoic with solid paleontological and sedimentological control, and most of the units have not experienced temperatures in excess of ~300 °C (Chamba: e.g., Webb et al., 2011; Leger et al., 2013; Srinagar and Pakistan Himalaya: e.g., DiPietro and Pogue, 2004). This stratigraphic continuity of low-grade rocks from the Indus suture zone to the Indian craton precludes the possibility of a hidden suture within the western Himalaya.

5.3. The Greater India Basin hypothesis rejected

We conclude that the Greater India Basin hypothesis is not supported by any solid piece of geological evidence. No "cryptic" suture can be located anywhere. No record of post-rift stratigraphic successions that should have been deposited along both GIB continental margins is preserved anywhere. The Tethys Himalaya was never separated from the rest of India. The collision between the Tethys Himalaya and Asia is the India-Asian collision.

6. The Late Cretaceous ophiolite obduction hypothesis

The Yarlung-Zangbo Ophiolite, marking the site of original contact between the Indian and Asian continental margins for over 2500 km, has been studied extensively (Gansser, 1980; Hébert et al., 2012; Dai et al., 2013; Wu et al., 2014a). One of the most controversial topics over the past three decades has been the idea that ophiolite complexes may have been obducted onto India in the Late Cretaceous, well before the onset of the India-Asia collision (Searle, 1986; Kelemen et al., 1988; Corfield et al., 2005; Garzanti et al., 2005; Garzanti and Hu, 2015).

Ophiolite obduction, defined as the emplacement of oceanic lithosphere onto a continent (Dewey, 1976) is best documented in the northern Oman mountains, where an up to 20 km-thick slab of ultramafic mantle and mafic crustal rocks were emplaced onto the Arabian continental margin in the Late Cretaceous (Coleman, 1981; Searle and Cox, 1999). The Oman model was applied to the Spongtang ophiolite of the western Himalaya (Searle et al., 1997; fig. 14 in Corfield et al., 2001), whereas in Tibet the possibility of Late Cretaceous to earliest Paleocene obduction onto India before collision with Asia was raised already by Sino-French Expeditions (Girardeau et al., 1985b), and by Ding et al. (2005) among others.

6.1. Debate on ophiolite obduction in the western Himalaya

The Spongtang Ophiolite exposed in the Zanskar Range formed originally in the mid-Jurassic at 177 ± 1 Ma and includes an arc-sequence dated at 88 ± 5 Ma (Reuber, 1986; Pedersen et al., 2001). At Late Cretaceous time, pelagic limestones were deposited all along the passive continental margin of India (Gaetani and Garzanti, 1991), followed by deposition of monotonous shelfal quartzose bioclastic siltstones derived from India in the south and lacking ophioliticlastic or volcaniclastic detritus altogether (Kangi La Formation; Gaetani and Garzanti, 1991). A widespread carbonate platform followed without break in the Paleocene (Nicora et al., 1987). As already pointed out by Kelemen et al. (1988) and Garzanti et al. (2005), there is no supporting evidence for pre-collisional obduction of the Spongtang Ophiolite onto India. Indian passive margin sedimentation did not record any obduction event and any erosion of ophiolitic material in the Late Cretaceous (Garzanti and Hu, 2015).

The mèlange underlying the Spongtang Ophiolite yielded early Eocene radiolaria and foraminifers, constraining emplacement of the Spongtang Ophiolite to the early Eocene (Colchen et al., 1987; Reuber et al., 1987). Moreover, lower Eocene Tethys Himalayan strata underlying the ophiolitic allochthon display upward-increasing, very low-grade metamorphism of mid-Eocene age (fig. 8 in Garzanti and Brignoli, 1989; Bonhomme and Garzanti, 1991), which confirms that thrusting of the Spongtang Ophiolite onto the Indian passive margin did not occur in the Late Cretaceous before the India/Asia collision, but took place when the Tethys Himalayan margin underthrust Asia in the early Paleogene.

6.2. Debate on ophiolite obduction in the central and eastern Himalaya

Pre-collisional, ophiolite-obduction events have been imagined to have taken place at different times during the Cretaceous also in southern Tibet. Abundant basalt, chert, serpentinite and pyroxene grains in the lower part of the Chongdui sandstone member, particularly in the Qunrang section, were interpreted as sourced from the Yarlung-Zangbo Ophiolite (Nicolas et al., 1981; Wang et al., 1999). Similar detritus exists in the overlying Ngamring turbidites (Dürr, 1996; Einsele et al., 1994; Wang et al., 1999; An et al., 2014), and detrital Cr-spinels from the Chongdui and Ngamring turbidites display identical geochemical features (Hu et al., 2014). Sources other than the Yarlung-Zangpo Ophiolite are possible, including ultramafic rocks of the Permian Sumdo Ophiolite in the southern Lhasa Block (Yang et al., 2009). Further studies are needed to establish the provenance of ophiolitic detritus found in mid-Cretaceous Xigaze forearc-basin strata.

Although ophiolitic detritus and low-Ti Cr-spinels found in Paleogene successions throughout the Himalaya are generally interpreted as derived from the Yarlung-Zangbo ophiolite (e.g., Ding et al., 2005), a distinction should be made between the western Himalaya and southern Tibet. Detrital Cr-spinels with very low TiO_2, high Al_2O_3 and low Cr#, similar to those from the Yarlung-Zangbo Ophiolite, do occur in lower to middle Eocene sandstones of the Zanskar Range, NW India and Pakistan, as in upper Eocene to lower Miocene sandstones of the Bengal Basin. In contrast, Cr-spinels from the Enba, Zhaguo, Sangdanlin, Zheya, Quxia and Jialazi formations of southern Tibet are characterized by high TiO_2, low Al_2O_3 and high Cr#, similar to those in Xigaze forearc strata (Hu et al., 2014). U-Pb age patterns of detrital zircons suggest provenance from the Lhasa Block for this latter group of Cr-spinels, whereas they indicate an Himalayan source for Eocene strata in NW India and mixed Himalayan and Lhasa Block sources for upper Eocene-lower Miocene successions of the Bengal basin (Hu et al., 2014). Ophiolites of the Indus-Yarlung suture zone started to be subaerially exposed and eroded at early Eocene times in the western Himalaya as recorded in the Subathu Formation (Garzanti et al., 1987; Critelli and Garzanti, 1994; Najman and Garzanti, 2000), whereas the Yarlung-Zangbo Ophiolite may have been exposed only at later times in the central-eastern Himalaya, as Cr-spinels similar to those found in the ophiolites of the Indus-Yarlung suture zone are not found in the Paleocene-Eocene Enba, Zhaguo, Sangdanlin, Zheya, Quxia and Jialazi formations of southern Tibet.

6.3. The Late Cretaceous ophiolite obduction hypothesis rejected

We conclude that the hypothesis of pre-collisional ophiolite obduction onto the Indian passive margin is not supported by geological evidence. On the contrary, conclusive stratigraphic evidence that the Yarlung-Zangbo Ophiolite represents the forearc basement of the Asian arc-trench system (An et al., 2014) proves that ophiolite emplacement onto India could not have taken place prior to closure of Neo-Tethys and onset of India-Asia collision. Initial India-Asia collision and ophiolite obduction (i.e., underthrusting of the Indian continental margin beneath the Transhimalayan subduction complex and forearc) are one and the same event.

7. Discussion

7.1. Timing of India-Asia collision onset

The timing of collision onset could be constrained directly only in the Sangdanlin section of southern Tibet by dating the earliest turbidites of Asian provenance deposited on top of the distal northern Indian margin (59 ± 1 Ma; DeCelles et al., 2014; Wu et al., 2014a; Hu et al., 2015a) (Figs. 5 and 14). This conclusion is based on the assumption that the Denggang turbidites were deposited onto the very edge of the Indian continental margin rather than a few hundreds of km away on oceanic crust, in which case the estimate for the age of initial continent-continent contact would be too old by a few Ma. Passive-margin turbidite fans may indeed extend for several hundreds of km into the ocean if associated with huge river systems draining vast continental interiors (e.g., Mississippi, Nile, Congo; Savoye et al., 2009), but for longer distances only where sourced from huge active orogenic belts (e.g., Indus, Ganga-Brahmaputra; Curray et al., 2003). In our case, however, India being a small continent and the Denggang Fm. thus unlikely to be associated with a river the size of the Nile or the Congo, the fan could hardly extend on oceanic crust for much more than a hundred kilometers. This would imply that the 59 ± 1 Ma estimate for initial continent-continent contact might be too old but not by more than 1 Ma, and thus still within the uncertainty of stratigraphic methods. Repeated intercalation of Indian-derived and Asian-derived turbidites indicates not only that the Sangdanlin Formation was deposited in a topographic depression that can be nothing else than the trench, but also that the two continental margins were not far apart. DeCelles et al. (2014 p.19) remarked: "*The only lithofacies in the Sangdanlin section that are reasonable candidates for oceanic deposits are the radiolarian chert and siliceous shale of the Sangdanlin Formation, which could have accumulated on the flexural forebulge or in the trench as the underlying Indian transitional/continental lithosphere entered the subduction zone. These observations support the interpretation that the Sangdanlin record resulted from initial collision of Indian and Asian continental lithosphere*". The intercalation of siliceous shales with quartzose turbiditic sandstones in the upper part of the underlying Denggang Formation (Fig. 6) concur in fact to indicate deposition on the possibly hyper-extended continental margin of "Greater India" rather than in the open ocean.

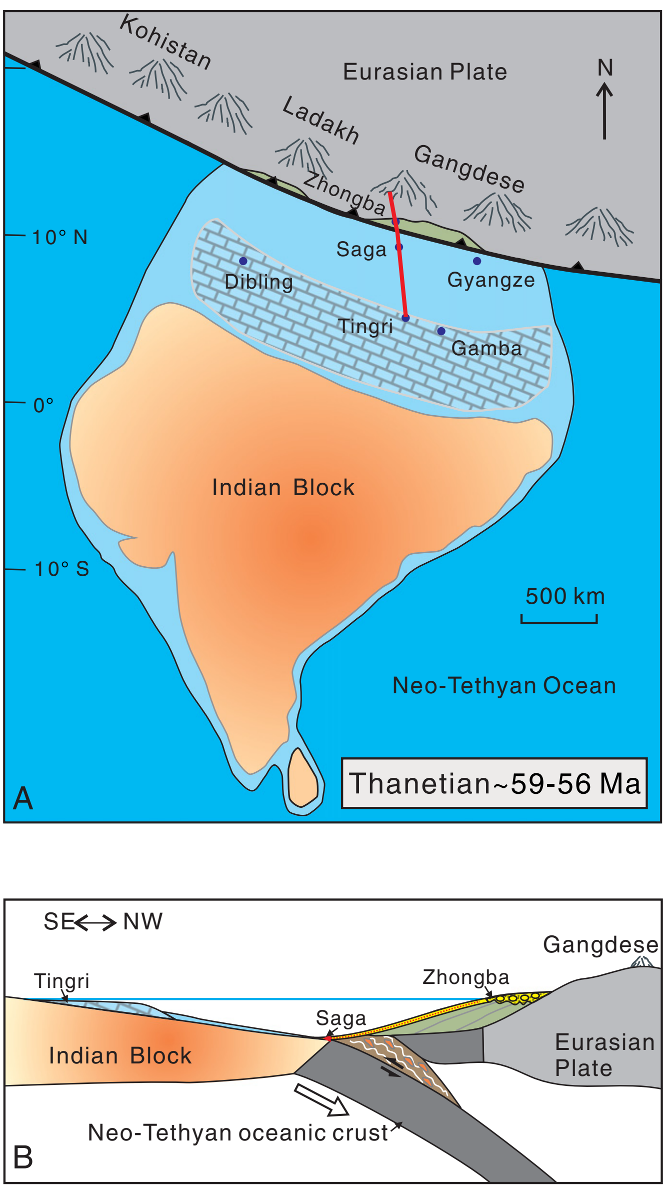

Fig. 14. (A) Cartoon showing full continent-continent contact between India and Asia during the Thanetian (between 59 and 56 Ma); (B) cross-section (location shown in Fig. 15A) of the Tibetan traverse at the instant of collision onset.

A second question concerns the relationship of the Sangdanlin section with surrounding units, as discussed in detail by DeCelles et al. (2014). The Sangdanlin section was mapped by Ding et al. (2005) as resting conformably upon Cretaceous strata of the northern Tethyan Himalayan zone, implying that it represents the Paleocene upper part of the Tethyan passive margin succession deposited on thinned Indian continental lithosphere. Instead, Wang et al. (2011) mapped the Sangdanlin section as completely bounded by faults and embedded within the mélange unit, implying that the Sangdanlin section might have been deposited oceanward of the Indian continental margin. In a third interpretation, the 1:250,000 geological map by Yang et al. (2003) includes the Sangdanlin section in the Zhongba-Gyangze thrust system, tectonically underlain and overlain by sedimentary rocks of Tethyan affinity. Following the interpretations of Ding et al. (2005) and Yang et al. (2003), DeCelles et al. (2014) placed the Sangdanlin section at the top of the Tethyan succession and therefore on the continental part of the Indian plate. Our further field observation confirmed that the Sangdanlin section is a tectonic slice within the mélange of the Yarlung Zangbo suture (Fig. 15), bounded by thrusts, as originally suggested by Wang et al. (2011). However, the mélange south to the Sangdanlin section formed largely after the India-Asia collision (An et al., 2016). In all cases the Sangdanlin section displays a continuous, north-dipping succession with well preserved stratigraphic relationships from base to top. Even though it may be locally affected by a few faults, in no way it can be interpreted as a mélange.

A third question concerns the provenance interpretation of the Enba + Zhaguo and Sangdanlin + Zheya formations, and whether they were sourced from the Gangdese arc or from an intraoceanic arc. Several lines of evidence support provenance from the Gangdese arc: 1) the Enba, Zhaguo, Sangdanlin, and Zheya formations all contain an abundance of felsic volcanic detritus (Fig. 16), whereas intraoceanic arcs are characterized by mafic to intermediate magmatism; 2) the Enba, Zhaguo, Sangdanlin, and Zheya formations contain abundant quartz grains that could not be derived from an intraoceanic arc (Fig. 16); 3) magmatism in the source areas had to be continuous from 200 to 60 Ma, as documented by U-Pb age spectra of detrital zircons from the Enba, Zhaguo, Sangdanlin, and Zheya formations, which is consistent with magmatism in the Gangdese arc prolonged from the Jurassic to the Paleocene (Chung et al., 2005; Ji et al., 2009; Zhu et al., 2011); 4) framework petrography, Cr-spinel-geochemistry, U-Pb age spectra and Hf isotopic ratios of detrital zircons (Fig. 16) characterizing the Sangdanlin and Zheya formations in Saga (Wang et al., 2011; Wu et al., 2014a; DeCelles et al., 2014), and the Enba and Zhaguo formations in Tingri-Gamba are very similar as those in the Quxia and Jialazi formations deposited on top of the Xigaze forearc basin, which was certainly derived from the Gangdese arc (Hu et al., 2016). We conclude that the

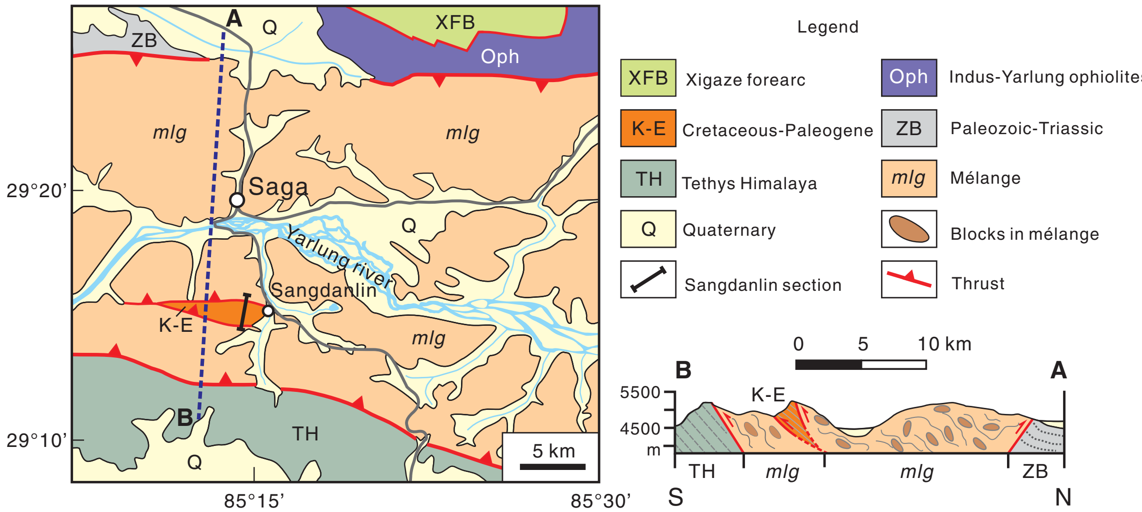

Fig. 15. Simplified geological map of the Sangdanlin area, showing that the Sangdanlin section is a tectonic slice bounded by faults within the mélange zone of the Yarlung Zangbo suture. A–B is the cross section across the Sangdanlin area.
(Modified after Wang et al., 2011).

same Gangdese source rocks provided detritus to the western Xigaze forearc basin as to the Enba, Zhaguo, Sangdanlin and Zheya formations.

7.2. Diachroneity of India-Asia collision onset

Accuracy and precision in dating early collisional stages has proven so far to be insufficient to detect any potential diachroneity from the western to the central-eastern Himalaya (DeCelles et al., 2004; Najman et al., 2005, 2010). The major unconformity interpreted as recording the passage of an orogenic wave is recorded at the Paleocene/Eocene boundary (~56 Ma) by the same inner-margin carbonate platform both in the Zanskar Range of the NW Himalaya (Garzanti et al., 1987) and in the Tingri and Gamba areas of southern Tibet (Hu et al., 2012; Zhang et al., 2012; Li et al., 2015b), indicating broadly synchronous post-collisional tectonic evolution along strike. The Paleocene-Eocene sedimentary successions of the NW Himalaya and of southern Tibet can be correlated interval by interval, from the Stumpata-Jidula quartzose sandstones of the Danian, to the Dibling-Zongpu carbonate platform of the Selandian-early Ypresian, to the Kong-Enba greenish mudrocks and sandstones of the late Ypresian/Lutetian?, to the overlying Chulung La-Zhaguo fluvio-deltaic red beds (Garzanti et al., 1987; Najman et al., 2010, 2016; Hu et al., 2012) (Fig. 17). The time-equivalence (within errors of few Ma) of main tectonic events, provenance changes, facies changes and sedimentary patterns documented all along the edge of the Indian continental margin throughout the Paleocene and early Eocene reduces the potential time lag between collision

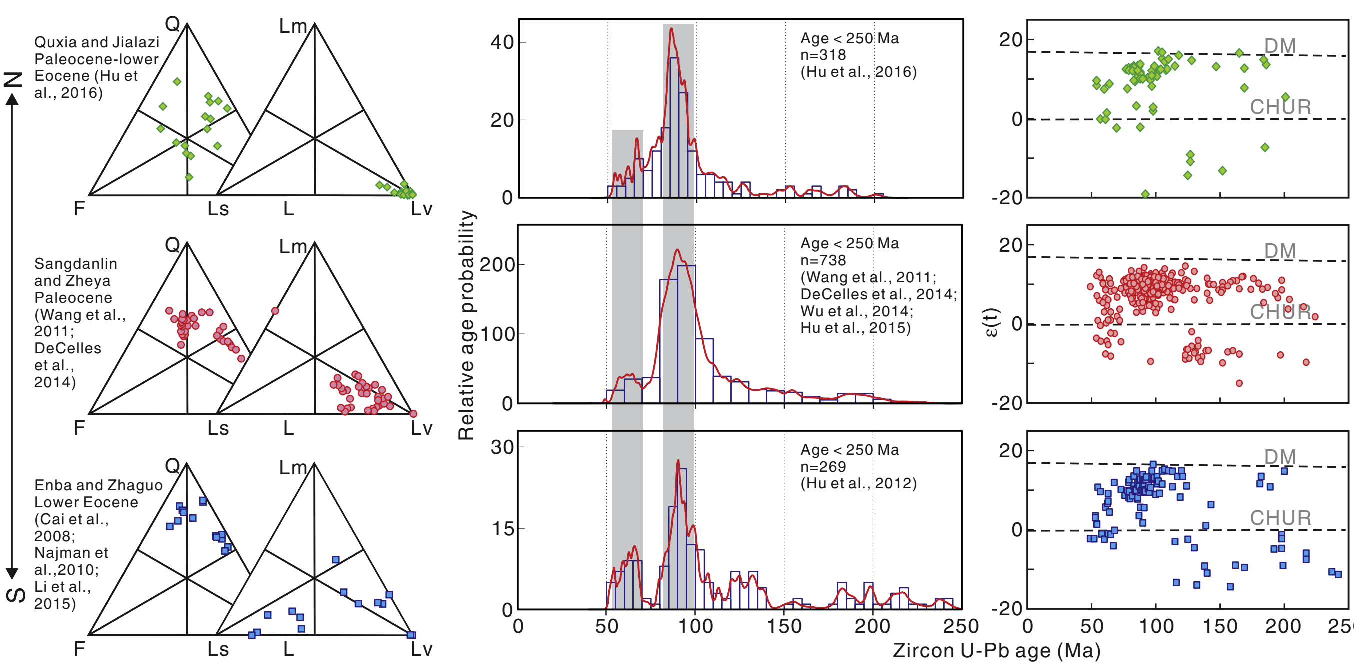

Fig. 16. Sandstone composition, age spectra and Hf isotopic signatures of detrital zircons in syncollisional basins (Quxia and Jialazi formations in Zhongba, Sangdanlin and Zheya formations in Saga, Enba and Zhaguo formations in Tingri). Left: QFL and LvLsLm diagrams; middle: Mesozoic-Paleogene ages of detrital zircons; right: Hf isotopic signatures of Mesozoic-Paleogene detrital zircons (locations shown in Fig. 15A). Data sources: Cai et al. (2008), Najman et al. (2010), Wang et al. (2011), DeCelles et al. (2014), Li et al. (2015b), and Hu et al. (2016).

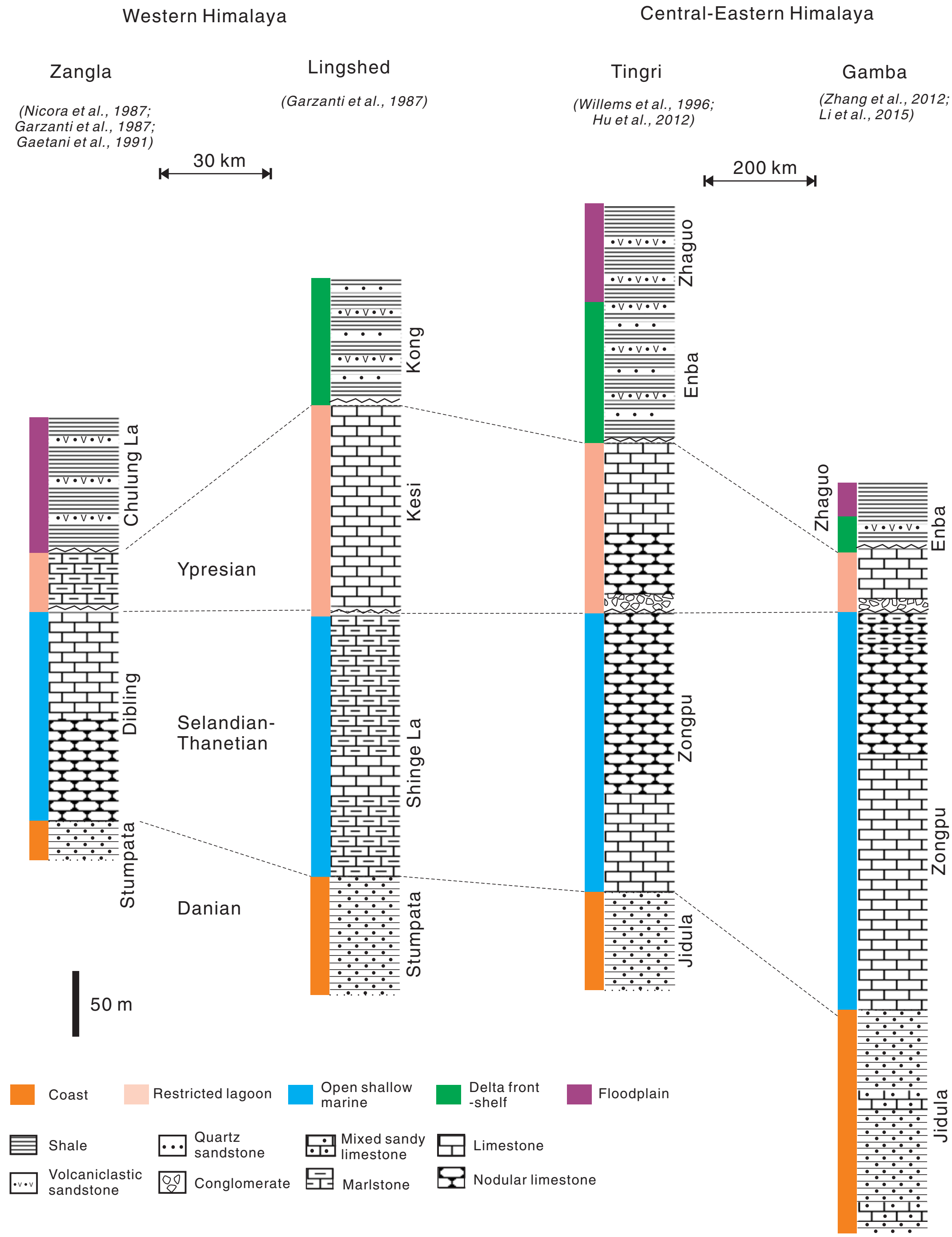

Fig. 17. Stratigraphic correlation of Paleocene-early Eocene events between the western Himalaya (Zangla and Lingshed) and the central-eastern Himalaya (Tingri and Gamba), showing similar, quasi-isochronous sedimentary evolution. Data sources: Nicora et al. (1987), Garzanti et al. (1987), Gaetani and Garzanti (1991), Willems et al. (1996), Hu et al. (2012), Zhang et al. (2012), and Li et al. (2015b).

onset in the western and central-eastern Himalaya to within the range of the uncertainties inherent to stratigraphic research (Fig. 17).

7.3. Assessment of methods used to constrain collision onset

A wide spectrum of disciplines has been used to unravel paleogeographic evolution during the early stages of the Himalayan orogeny. Not all of them, however, are equally effective, or yield truly independent information. Most offer only indirect if not circumstantial clues that can be interpreted in different alternative ways. The cessation of marine sedimentation, provenance changes in inner margin sections, onset of tectonic deformation, or the age of HP-UHP metamorphism can only provide a minimum estimate for collision onset (Fig. 18). Insights obtained from changes in magmatic activity, convergence rate, remagnetization, changes in basin type, or events of post-collisional intramontane sedimentation are ambiguous, and their interpretation might be influenced by the pre-conceived views of the interpreter.

The sedimentary record represents the best ally in the reconstruction of collision events leading to orogenic growth. Interpreting the magmatic, metamorphic or paleomagnetic record generally requires a greater number of assumptions and consequently suffers from a larger variety of uncertainties. Although all approaches potentially provide

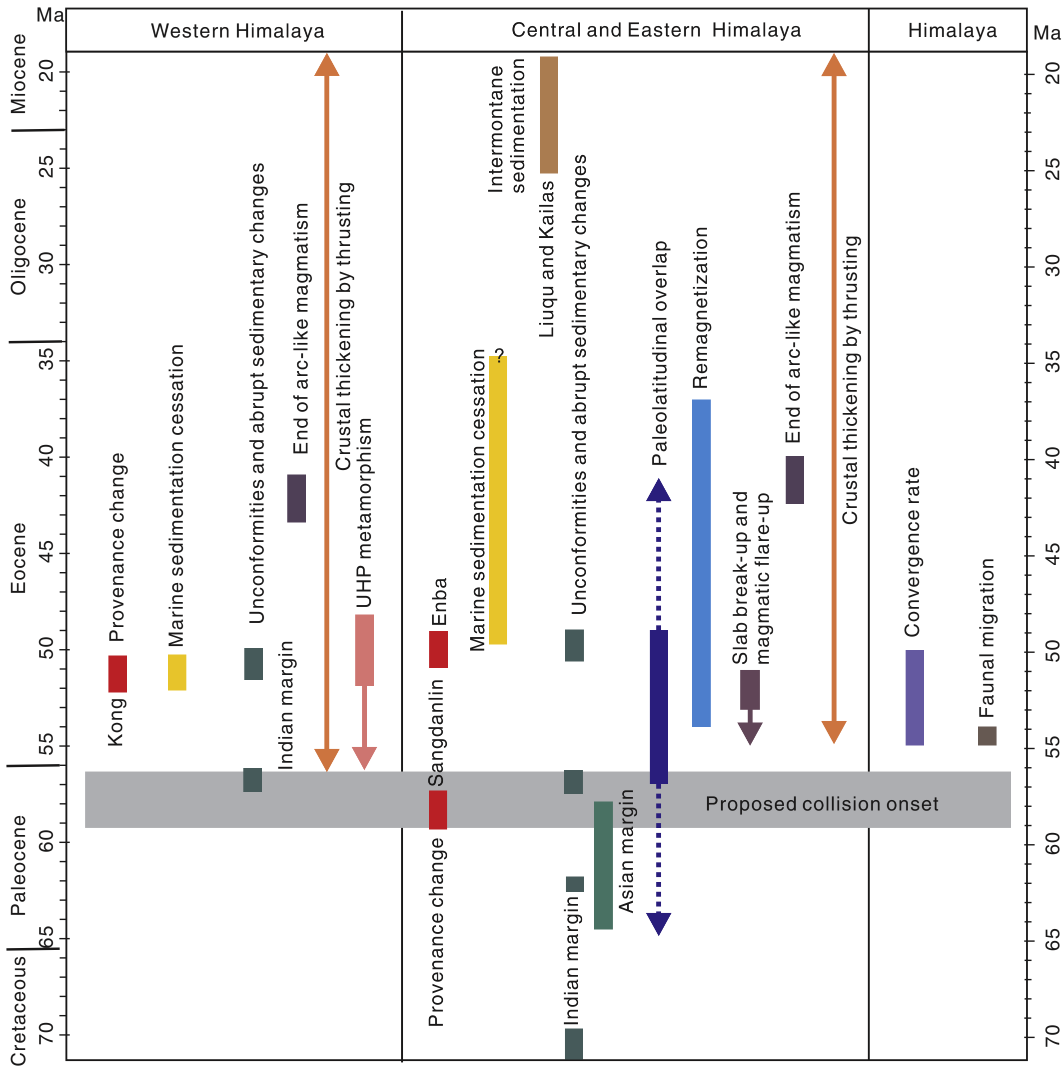

Fig. 18. A summary of various methods used to constrain India-Asia collision onset. Collision onset can be dated directly from the sedimentary record of the distal Indian continental rise (e.g., Sangdanlin section), as documented by a radical change in sediment provenance. The cessation of marine sedimentation, provenance changes documented in inner margin sections, or HP-UHP metamorphism can only provide a minimum estimate for collision onset. Preserved records of intermontane sedimentation extend to 40 Ma after collision onset.

indispensable elements to fit in the paleotectonic puzzle, our advice is to have both feet firmly grounded in field observations before giving birth to imaginary collision scenarios that contradict basic geological facts (Gansser, 1991).

Recommended approaches that proved to be particularly useful to accurately reconstruct the chronology of continental collision and the subsequent syn-collisional evolution of mountain belts grown during continental subduction include the following:

(1) in the distal margin of the lower plate, the stratigraphically-dated arrival at the trench documented by interfingering turbidites shed from the opposite upper-plate margin provides a most accurate approximation to date collision onset;
(2) in the inner margin of the lower plate, stratigraphic dating of a major unconformity associated with tectonic uplift, followed up-section by the transition from marine to fluvio-deltaic sediments derived from the opposite upper-plate margin, provides a robust minimum-age constraint for collision onset and a record of early syn-collisional evolution;
(3) the exchanges of shallow- marine and terrestrial faunas documented by the detailed paleontological study of stratigraphic successions on both lower and upper plates may provide robust minimum-age constraints for the disappearance of deep oceanic troughs first, and for the disappearance of shallow seaways next;
(4) the age of HP-UHP metamorphism of continental protoliths exposed close to the suture zone may provide a robust minimum-age constraint for collision onset.

8. Conclusion and perspectives

The timing of initial collision between India and Asia has remained grossly and windingly controversial for half a century, partly because different approaches often implied different criteria to define collision onset, and partly because different researchers put their faith in different proxies, sticking to evidence provided by their own approach while often dismissing too lightly evidence provided by different techniques or other research groups. The modern trend toward a decay of the facts/theories ratio in Himalayan geological research was long foreseen by Augusto Gansser (1991 p.35), who nostalgically noted "*even in the presentation of facts … we note theoretical interpretations hardly representing nature any more*". The present review has attempted to detangle the much confused issue of India-Asia collision timing, and to show how the clarity of facts has often been dimmed by a whirl of theories not invariably founded on solid rock. To this goal we have reviewed the diverse paths chosen to attack and solve the problem, highlighting their rationales, results, and shortcomings.

(1) Stratigraphy offers the best way we dispose of to look directly into collision events. Other approaches based on the interpretation of magmatic, metamorphic or paleomagnetic records provide fundamental additional constraints. All methods suffer from uncertainties, but indirect methods suffer far more because they are based on several assumptions (and sometimes on assumptions based on other assumptions).
(2) Collision onset is dated directly at 59 ± 1 Ma in the Tibetan traverse by combined biostratigraphy and zircon chronostratigraphy of turbidites derived from India, interfingering stratigraphically with turbidites derived from Asia in the deep-water Sangdanlin Formation deposited in the Transhimalayan trench. The unconformities observed both on the inner Indian margin and on the Asian margin support collision onset before the end of the Paleocene (>56 Ma).
(3) The correlative Paleocene-Eocene shallow-marine to fluvio-deltaic successions deposited in the western Himalaya and in the central-eastern Himalaya favor quasi-synchronous collision within the limits of stratigraphic precision.
(4) The three widely cited hypotheses of Late Cretaceous ophiolite obduction, Paleogene arc-continent collision, and Greater India Basin are rejected after discussing point by point the supporting and contradicting geological evidence.
(5) No more complex scenario than the one simply involving the passive continental margin of India and the active continental margin of Asia is needed to explain the geological evolution of the nascent Himalaya. The collision between the Tethys Himalayan passive margin and the Asian active margin and the collision between India and Asia are one and the same event. The emplacement of the Yarlung Zangbo forearc ophiolites onto the Tethys Himalaya and underthrusting of the Tethys Himalaya beneath the Asian arc-trench system (i.e., India-Asia collision) are one and the same event.

The following suggestions are made for the benefit of future studies aimed at obtaining observable geological facts.

(1) More work is needed on the Chongdui Formation - especially on the contacts between Chongdui cherts and turbidites, and between Chongdui and Ngamring turbidites – in order to conclusively define the nature of the Yarlung Zangbo Ophiolite and the early stages of sedimentation in the Xigaze forearc basin. More work is needed also on the Xialu Chert Formation of Jurassic-Early Cretaceous age exposed south of the Yarlung Zangbo Ophiolite (Matsuoka et al., 2002; Ziabrev et al., 2004), which, unlike the chert member of the Chongdui Formation, has undergone strong tectonic deformation and may represent the sedimentary cover of Neo-Tethyan oceanic crust (Burg and Chen, 1984).
(2) More detailed and integrated biostratigraphic studies are needed on key intervals - including the Jiachala Formation and the Enba and Zhaguo formations in southern Tibet - in order to define more accurately the paleogeography of initial India-Asia collision and the subsequent early syn-collisional stages. Dating of Paleogene strata still depends heavily on biostratigraphic work, valuably supported by detrital-zircon chronostratigraphy most useful in continental deposits.
(3) Robust paleomagnetic data from Cretaceous to Eocene Tethys Himalayan strata are urgently required to test the potential motion of the Tethys Himalaya relative to India. Future paleomagnetic studies should also focus on the Late Cretaceous to Paleocene latitude of the Lhasa Block, which is still not fully constrained and controversial. Insidious remagnetization and inclination shallowing should be thoroughly tested and corrected before the acquired paleomagnetic directions are used for tectonic reconstruction.
(4) Detailed structural work is needed to assess the "missing" amount of intracontinental shortening accurately. The early record of crustal shortening of the northern Indian margin could have been either subducted beneath the Lhasa block along the Indus-Yarlung suture zone or eroded away during later orogenic evolution, or simply not recorded by any studies to date. The key subject of study would be the thin-skinned fold-thrust deformation of the Tethys Himalayan domain.
(5) Mélange units such as the Xiukang mélange or the Zongzhuo Formation have been investigated insufficiently. They can provide essential information on erosional, accretionary and deformation processes that took place before, during, and after collision.
(6) Extensive stratigraphic and provenance work has been carried out on syn-collisional deposits within and at the front of the nascent Himalayan belt. However, questions concerning their relationships and subsidence mechanisms remain unanswered because the two depositional systems are not connected and separated by Greater Himalayan metamorphic rocks. Whereas foreland-basin stratigraphy during the Neogene is reasonably well reconstructed, foreland-basin evolution during the Paleogene early collisional stages, and specifically the widespread occurrence of the late Eocene-Oligocene unconformity, are still poorly understood.

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Appendix A. Supplementary data

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