New insights into the timing of the India–Asia collision from the Paleogene Quxia and Jialazi formations of the Xigaze forearc basin, South Tibet

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A B S T R A C T

The Xigaze forearc basin provides information on subduction evolution and magmatic growth of the Gangdese arc as well as on the India–Asia continental collision. Recently obtained sedimentological, biostratigraphic, petrographic, geochemical and geochronological data on Cretaceous to Paleogene strata in the Cuojiangding area (Zhongba county, south Tibet) shed new light on the tectonic evolution of the southern margin of the Lhasa Block during closure of Neotethys and initial collision with India. The uppermost Cretaceous Padana and Qubeiya formations, deposited in deltaic to inner shelf environments, and representing the final filling of the Xigaze forearc basin, were unconformably overlain by the Quxia and Jialazi formations, deposited in fan-delta environments during the Paleocene/earliest Eocene. Petrographic data and U–Pb ages of detrital zircons document the progressive unroofing of the Gangdese arc, which remained the dominant source of detritus throughout the Late Cretaceous to Paleogene. Detrital Cr-spinels in the Quxia and Jialazi formations are geochemically similar to those in Cretaceous Xigaze forearc strata but different from those hosted in Yarlung Zangbo ophiolites, suggesting that the latter were not exposed to erosion in the considered time window. Sandstone petrography, Cr-spinel-geochemistry, U–Pb age spectra and Hf isotopic ratios of detrital zircons in the Quxia and Jialazi formations match those in Paleogene sediments deposited on the distal (Sangdanlin and Zhaya formations) and proximal Indian margin (Enba and Zhaguo formations), suggesting that the Quxia and Jialazi formations documents syncollisional fan-deltas deposited on top of the nascent Himalayan orogenic belt. In this scenario, the onset of the India–Asia collision predates deposition of the Quxia and Jialazi formations and is thus constrained as younger than 66 Ma and older than 58 Ma.

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1. Introduction

One of most hotly debated topics concerning the Himalayan orogeny is the timing of the initial India–Asia collision (e.g., Aitchison et al., 2007; Garzanti, 2008; Najman et al., 2010; Hu et al., 2012; DeCelles et al., 2014; Wu et al., 2014). Most paleomagnetic studies from both sides of the Yarlung Zangbo suture zone (YZSZ) in southern Tibet suggest that the initial contact between India and Asia took place around 55 ± 5 Ma (e.g. Huang et al., 2010; Najman et al., 2010; Meng et al., 2012; Sun et al., 2012; Yang et al., 2015 and references therein). Cessation of marine sedimentation and first arrival of Asian (Gangdese) detritus on the Indian margin is documented in both Tethys and Saga areas of southern Tibet to have occurred in the Early Eocene (Blondeau et al., 1986; Zhu et al., 2005; Najman et al., 2010; Wang et al. 2011; Hu et al., 2012), which confirms a minimum age of ~50 Ma for the India–Asia initial collision, although this age is debated by other researchers (e.g., Wang et al., 2002; Aitchison et al., 2007). Numerous studies on collisional orogens indicate that marine sedimentation may continue for several to tens of millions of years after the onset of continental subduction (e.g., Miall, 1995; Sinclair, 1997; Allen and Allen, 2013). The India–Asia collision may thus have initiated significantly earlier than the cessation of marine facies in the Tethyan Himalaya (~50 Ma).

One of the consequences of continental collision is the development of new sedimentary basins (e.g., Beaumont, 1981; Miall, 1995; Allen and Allen, 2013). Early foreland-basin successions have been thoroughly studied in south of the Himalayan range in Pakistan, India and Nepal (DeCelles et al., 1998; Najman and Garzanti, 2000; Singh, 2003; Najman et al., 2004), as well as in the Tethys Himalaya of South Tibet (Saga and Gyange: Ding, 2003; G.B Li et al., 2005; Wang et al., 2011; Wu et al., 2014; Tiangri and Gamba: Hu et al., 2012; Q.H. Zhang et al., 2012; Li et al., in press) (Fig. 1a). Similar development of syncollisional basins is documented also in the NW Himalaya, where evidence for
development of a flexural bulge in the outer Indian margin has been constrained biostratigraphically at the Paleocene/Eocene boundary (~56 Ma; Garzanti et al., 1987; Garzanti, 2008). Such an age for collision onset corresponds closely with the onset of India–Asia faunal exchange (occurred at ~54 Ma according to Clementz et al., 2011), with the possible initiation of deformation of the Indian margin (Ratschbacher et al., 1994; Ding et al., 2005), and with the age of NW Himalayan eclogites (Guillot et al., 2003; Leech et al., 2005).

Aitchison and his group (e.g., Abrajевичt et al., 2005; Aitchison et al., 2007) proposed that India collided first with an intraoceanic arc at ~55 Ma, and eventually only at ~35 Ma with Asia, Asian-derived detritus in Paleogene Tethyan Himalayan sandstones (Enba Formation in Tingri and Sangdanlin Formation in Saga) (Zhu et al., 2005; Najman et al., 2010; Wang et al., 2011; Hu et al., 2012; J. Zhang et al., 2012; Q.H. Zhang et al., 2012; DeCelles et al., 2014; Wu et al., 2014), however, indicates that at ~55 Ma India collided with the Lhasa Block earlier attached to Eurasia, thus ruling out the idea of a double Cenozoic collision. No geological evidence of oceanic suturing and collision later than 55 Ma has ever been reported either in the north between the collision along the Asian active margin (Fig. 1a). In this study, we combine sedimentological and biostratigraphic data on the Upper Cretaceous to lowermost Eocene succession of the Cuojiangding area, with sandstone petrography, detrital zircon U–Pb geochronology, and detrital Cr-spinel geochemistry to reconstruct the paleotectonic evolution of the southern margin of Asia during transition from Neo-Tethyan subduction to the earliest stages of the India–Asia collision.

Fig. 1. Simplified tectonic map of the Himalaya (a; after Gansser, 1964). Inset b): geological sketch map of the Zhongba area, southern Tibet (after the 1:1,500,000 geologic map of Pan et al., 2004). Cross section marked A–A' outlines the NW–SE transect from Cuojiangding to Sangdanlin and Tingri shown in Figs. 10 and 11.
2. Geological setting

2.1. The Himalaya

The Himalayan orogenic belt is traditionally thought to consist of a series of east/west trending litho-tectonic units (Fig. 1; Gansser, 1964), described here below from north to south.

1. The Lhasa Block north of the Gangdese arc includes Paleozoic to Cretaceous strata with volcanic intervals and common Jurassic to Cretaceous igneous rocks (Kapp et al., 2005; Zhu et al., 2011). Significant thrusting prior to the India–Asia collision (Murphy et al., 1997; Kapp et al., 2007) and associated retroarc-foreland-basin development were inferred to have taken place during the Late Cretaceous (Leier et al., 2007a; Sun et al., 2015). The Gangdese magmatic arc (southern Lhasa subterrane) is composed of Late Triassic to Paleogene granitoid batholiths (Chung et al., 2005; Chu et al., 2006; Wen et al., 2008; Ji et al., 2009; Zhu et al., 2011), intruding Paleozoic and Mesozoic strata, and unconformably overlain by the Paleogene Linzizong volcanic succession (Mo et al., 2008; Lee et al., 2009; Zhu et al., 2013; Zhang et al., 2014).

2. The Xigaze forearc basin was filled by turbiditic sandstones and interbedded mudrocks during the Late Cretaceous (Leier et al., 2007a; Sun et al., 2015). The Gangdese magmatic arc (southern Lhasa subterrane) is composed of Late Triassic to Paleogene granitoid batholiths (Chung et al., 2005; Chu et al., 2006; Wen et al., 2008; Ji et al., 2009; Zhu et al., 2011), intruding Paleozoic and Mesozoic strata, and unconformably overlain by the Paleogene Linzizong volcanic succession (Mo et al., 2008; Lee et al., 2009; Zhu et al., 2013; Zhang et al., 2014).

3. The Yarlung Zangbo suture zone marks the contact between India and Asia, and includes ophiolites and tectonic mélangé with serpentinite or shale matrix (Cai et al., 2012; Chan et al., 2015; Xu et al., 2015).

4. The Tethyan Himalayan sedimentary sequence is generally subdivided into southern and northern zones, separated by the Gyirong–Kangmar thrust (Ratschbacher et al., 1994) and representing proximal and distal northern Indian continental-margin depozones, respectively. The southern Tethyan Himalaya is characterized by shallow-water shelf carbonates and terrigenous Paleozoic to Eocene strata (e.g. Jadoul et al., 1998; Garzanti, 1999; Hu et al., 2010), whereas the northern Tethyan Himalaya is dominated by deeper-water Mesozoic to Paleogene outer shelf, continental slope and rise deposits (X.H. Li et al., 2005).

2.2. The Xigaze forearc basin

The Xigaze forearc basin is exposed in southern Tibet over a length of 510 km with a maximum width of 22 km. The first modern sedimentologic, biostratigraphic, petrologic and tectonic studies were carried out in the 1980s, when the Sangzugang, Chongdui, Ngamring, and Padana formations were established (Liu et al., 1988; Fig. 3). In the 1990s, sedimentological and provenance studies documented the predominance of submarine-fan deposits, fed mostly from the Gangdese arc but also partly from a carbonate platform and the Lhasa terrane to...
the north (Einsele et al., 1994; Dürr, 1996; Wang et al., 1999). Biostratigraphic studies dated the Ngamring Formation at the late Albian–Coniacian, and subdivided it into five megasequences (Wan et al., 1998; Wang et al., 2012). Extensive detrital-zircon U–Pb–Hf data confirmed provenance from the Gangdese arc (Wu et al., 2010; Aitchison et al., 2011; Orme et al., 2015). Most recently, An et al. (2014) documented in detail how sedimentation in the forearc basin began with abyssal cherts of late Barremian–Aptian age in the south (Chongdui Fm.) and reefal limestones of late Aptian–earliest Albian age in the north (Sangzugang Fm.), and how the thick overlying turbidites of the Ngamring Fm. are capped by shelfal, deltaic and eventually fluvial sediments of the Padana Formation, documenting the final filling of the basin by Campanian time.

2.3. The Cuojiangding section

The studied section is located near the glacier lake of Cuojiangding, and geologically belongs to the Xigaze forearc unit (Qian et al., 1982;...
Liu et al., 1988; Ding et al., 2005) (Fig. 1b). The Quxia and Jialazi formations, first reported by Qian et al. (1982), are exposed in a syncline separated by thrusts from the Gangdese batholith and Linzizong volcanic rocks to the north, and from the Zhongba ophiolitic mélange to the south (Fig. 2). Its lithostratigraphy and sedimentary petrology was studied by Liu et al. (1988) and Jia et al. (2005), whereas paleontological studies were carried out on larger benthic foraminifers (Liu, 1988; Wan et al., 2001), ammonites (Sun and Wang, 2001) and pollens (Li et al., 2008). Ding et al. (2005) presented new data on the geological structure, geochronology, and geochemistry of detrital Cr-spinels.

3. Analytical methods

The Cuojiangding section was logged and described in detail during the 2009 and 2010 field seasons. More than 150 samples were collected for laboratory study. Thin-sections were prepared for all samples, and
23 least altered and coarser-grained sandstones were selected for petrographic analysis. About 400 points were counted on each sample following the Gazzi-Dickinson point-counting method (Ingersoll et al., 1984). Sandstones were classified according to their main components exceeding 10% QFL (quartz, feldspars and lithics) (e.g., in a lithofeldspatho-quartzite sand Q > F > L > 10% QFL).

Heavy minerals were separated from sandstone samples by elutriation and magnetic separation. Zircon grains were handpicked randomly, mounted in epoxy resin and polished for analyses. U–Pb dating of detrital zircons was carried out by laser ablation–inductively coupled plasma-mass spectrometry (LA-ICP-MS), following Jackson et al. (2004). Various beam diameters (18 μm to 35 μm) were used depending on the size of zircon grains. We obtained for Mud Tank Zircon a weighted 206Pb/238U age of 7244 ± 8.0 Ma (2σ, n = 29), in agreement with the recommended value (TIMS age = ±732 ± 5 Ma: Black and Gulson, 1978). GLITTER 4.4 was used for calculations of analytical results and relevant isotopic ratios (VanderAchterbergh et al., 2001). Age calculations and plotting of concordia diagrams were performed by Isoplot 3.75 (Ludwig, 2012). Zircons > 200 Ma with discordance <10% and <200 Ma with discordance less than 30% were used. Zircon-age statistics was carried out using the Age Pick program of the University of Arizona (http://www.geo.arizona.edu/alt/Analysis%20Tools.htm).

Because 206Pb/238U ages are more precise for younger zircons whereas 207Pb/206Pb ages are more precise for older zircons, we based our interpretations on 206Pb/238U ages for grains younger than 1000 Ma and on 207Pb/206Pb ages for grains older than 1000 Ma. In order to constrain the maximum depositional ages for each sample, we used five methods discussed by Dickinson and Gehrels (2009): (a) youngest single grain age (YSG); (b) youngest graphical age peak controlled by more than one grain age (YPP); (c) mean age of the youngest two or more grains that overlap in age at 1σ (YC1); (d) mean age of the youngest three or more grains that overlap in age at 2σ (YC2); (e) youngest detrital zircon age (YDZ) generated by Isoplot 3.75 (Ludwig, 2012).

Younger U–Pb age zircons (<250 Ma from the Quxia and Jalialzi formations and ~70 Ma zircons from the Sangdanlin and Zheya formations); U–Pb data provided in Wang et al. (2011) were selected for Lu–Hf isotopic analyses, using a Thermo Scientific Neptune Plus MC–ICP–MS coupled to a New Wave UP193 solid-state laser ablation system following the procedure described by Griffin et al. (2000). Hf isotopic compositions were determined with a laser beam diameter of 35 μm (laser repetition rate 8 Hz) or 25 μm (laser repetition rate 7 Hz), with an energy of 15.5 J/cm². The complete data set is given in Appendix Table 5. To calculate model ages (TDM) and epsilon Hf, we have assumed depleted mantle with 176Hf/177Hf = 0.282525 and 176Lu/177Hf = 0.0384, and chondrite with 176Hf/177Hf = 0.282772 and 176Lu/177Hf = 0.0332 (Griffin et al., 2000). The adopted decay constant for 176Lu is 1.867 × 10⁻¹¹ per year (Söderlund et al., 2004).

Cr-spinel composition was determined using a JEOL JXA-8100M electron microprobe with accelerating voltage 15 kV, beam current 20 nA, beam diameter 1 μm, and ZAF correction model. Detecting time was 10 s for Al, Fe and Mg, 20 s for Ti, Mn, Cr, V and Ni, and 30 s for Zr. Detection limits were ~200 ppm for all elements. All Fe was expressed as FeO, and the ferric iron content of each analysis was calculated following Barnes and Roeder (2001). All analyses were conducted at State Key Laboratory for Mineral Deposits Research, Nanjing University. Data are listed in the Appendix Tables 2 to 5.

4. Results

4.1. Stratigraphy and sedimentology

In the Cuojiangding area (Figs. 1b and 2), the Upper Cretaceous Padana Formation and the overlying Qubeiya Formation are unconformably followed by the Paleogene Quxia and Jalialzi formations (Liu et al., 1988; Ding et al., 2005) (Fig. 3). Correlation between relative and absolute ages is according to the Gradstein et al. (2012) timescale throughout the article.

4.1.1. Padana Formation

The Padana Formation in the Cuojiangding area is ~640 m-thick and compares well to the type area in the Padana valley (SE of Sangsang, Angren county; Liu et al., 1988; An et al., 2012). It mainly consists of gray sandstone interbedded with mainly grayish to reddish shale with common carbonate concretions (Fig. 4a). Sedimentary features include parallel to megaripple lamination, graded bedding, caliche soil profiles, coquina beds, mud intrasets and bioturbation (particularly abundant in red shales), indicating upward progradation from delta-front to delta-plain facies (Fig. 3). The ammonites reported by Sun and Wang (2001) include Manambolites cuajangdingensis, M. pivaeteae, Coahuilites zhongbaensis sp. nov., Libycoceras tibeticum sp. nov., L. zhongbaense sp. nov., indicating a late Campanian age. The youngest detrital zircon ages of the 09CJD03 sample from the top of the Padana Formation are 74 ± 1 Ma, 75 ± 1 Ma, 76 ± 1 Ma and 79 ± 1 Ma with discordances of 8%, 13%, 0% and 1%, respectively. YCC1(3+) and YDZ ages are also close to 74 Ma (Table 1), consistent with the late Campanian depositional age of the unit.

4.1.2. Qubeiya Formation

The Qubeiya Formation (~200 m thick in the Quxia A section; Fig. 3) conformably overlies the Padana Formation. It consists of yellowish-gray, fine-grained sandy wackestones intercalated with fine-grained sandstones. Abundant fossils include larger benthic foraminifera (mainly Lentinorhodontoidea spp.) (Fig. 4c, f), bivalves, ammonites, gastropods and crinoids, indicating Maastrichtian age and deposition on the inner shelf (Liu et al., 1988; Sun and Wang, 2001; Wan et al., 2001).

In this study we retrieved rich benthic foraminiferal assemblages including Lentinorhodontoidea minor, L. blanfordi, Orbitoides medius, O. vredenburgi and Omphaloclyscus spp. (Fig. 5). These Orbitoides lived in shallow tropical and subtropical seas with high light intensity (Boudagher-Fadel, 2008). The samples with larger and more robust...
Fig. 5. Stratigraphic distribution of larger benthic foraminifera in the Cuojiangding succession. SBZ: Shallow Benthic Zones of Serra-Kiel et al., 1998. Time scale after Gradstein et al. (2012).
orbitoids commonly contain gastropods and only rare fragments of algae and corals (Hottinger, 1997). The presence of L. minor with quadriserial peri-embryonic arrangement and small number of adaxialy chambers, in association with Orbitoides medius and Omphalocyclus, indicates a late Maastrichtian age (Fig. 5).

4.1.3. Quxia formation

The ~15 m-thick lower part of the unit consists of grayish to reddish sandstones (Fig. 3). Trough cross-lamination in sandstone beds and scoured basal contact suggest fluvial-bar environments, whereas the interbedded sandstones are interpreted as flood-plain deposits. The ~90 m-thick middle and upper parts of the unit consist of conglomerates, sandstones and shales arranged in five fining-upward cycles (Fig. 3). Conglomerate clasts are mostly 2–10 cm in diameter and well rounded, but cobbles up to 20 cm in size occur in the lower cycle (Fig. 4d). Pebbles of volcanic and volcanioclastic rocks (45% andesite and rhyolite; 15% tuffs) prevail over granitoid (15%), chert (15%), limestone (5%) and sandstone (5%) clasts. Sedimentary structures (graded bedding, imbrications, tabular cross-bedding) indicate deposition in a fan delta, with massive conglomerates interpreted as debris-flow deposits. Age-diagnostic fossils are lacking. The youngest detrital zircon U–Pb ages are 65.5 ± 0.8 Ma and 66 ± 1 Ma (2 grains), with discordance of 4% and 2.9%, respectively. The YDZ, YSG and YC10 ages are also –66 Ma (Table 1), which indicates the early Paleocene as the maximum depositional age of the Quxia Formation.

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), but detailed regional geological mapping including nearby areas such as that northeast of the Lopu Kangri Mountain (Orme et al., 2015) is needed to solve the issue.

4.1.4. Jialazi Formation

This ~240 m-thick unit conformably overlies the Quxia Formation and mainly consists of sandy foraminiferal limestones and interbedded sandstones (Figs. 3 and 4b). Sedimentary structures including graded bedding and oblique lamination indicate deposition in shallow-marine, distal fan-delta environments. Abundant fossils include larger benthic foraminifera, bivalves, gastropods and corals of Paleocene to early Eocene age (Liu et al., 1988; Wan et al., 2001).

In the lower Jialazi Formation (unit 15 of the Quxia section), larger benthic foraminiferal assemblages including Daviesina langhiami, Lockhartia haimei, L. cushmani, L. conditi, Ranikothalia bermudezi and Smoutina cruyci indicate the late Paleocene (early Thanetian, late SBZ3 zone of Serra-Kiel et al., 1998). The foraminiferal assemblages that we retrieved from the overlying units 16–18 are dominated by the larger benthic foraminifera Miscellanea yvettae, M. dukhani, Miscellanea sp. A, Ranikothalia sindensis, R. sahnii (Fig. 5), associated with the planktonic foraminifera Pseudomaniella ehrenbergi, Morozovella angulata, Acrarinina sp., Globanomalina planoconica, Globigerina sp. and Subbotina sp. This varied fauna indicates a latest Paleocene age (late Thanetian, SBZ 4 zone, late P4 zone; 57–56 Ma).

In the upper Jialazi Formation (units 24–25 of the Quxia section and the Jialazi section), larger benthic foraminifera include Assilina leymieriei, A. dandotica, A. prica, Miscellanea miscella, Nummulites deserti, Operculina canallfera, Ranikothalia sindensis, R. nutalli and R. sibetica, indicating an earliest Eocene age (Ypresian; SBZ5 zone; early P5 zone; 56–55 Ma) (Fig. 5). Fragments of rodophyte algae, Lithothamnium and the coralline Distichophas biserialis are widespread, indicating shallow-reef to fore-reef environments (BouDagher-Fadel, 2008).

Two rhyolite tuffs in the lower Jialazi Formation (unit 19 of the Quxia section) mostly consist (~80%) of partially to completely devitrified glass shards with needle-like and irregularly polygonal shapes, with ~5% quartz crystals. Zircons from these tuffs yielded weighted mean ages for the youngest age peak of 54.9 ± 0.7 Ma (n = 28, sample 09QX-B19) and 55.7 ± 0.5 Ma (n = 19, sample 10QX-B03) (Fig. 6).

4.2. Sandstone petrography

Sandstones of the Padana Formation are quartz–lithic volcanioclastic (average composition Q:F:L = 21:8:71, Fig. 4e, Appendix Table 2). Quartz grains are mainly monocrystalline and subangular to subrounded. Lithic fragments are mainly volcanic (Fig. 7b); K-feldspars and plagioclase are equally abundant.

Sandstones of the Quxia Formation are feldspatho–quartz–lithic volcanioclastic (average composition Q:F:L = 33:28:39, Fig. 4g, Appendix Table 2). Quartz grains are mainly monocrystalline. The cement is mainly calcite. Volcanic rock fragments are mainly andesitic and rhyolitic (Fig. 7b), and both K-feldspar and plagioclase occur.

Sandstones of the Jialazi Formation are feldspatho–litho–quartzose volcanioclastic (average composition Q:F:L = 47:22:31, Fig. 4h, Appendix Table 2). Quartz grains are monocrystalline and angular to subrounded. Volcanic rock fragments are mainly andesitic and rhyolitic, and both K-feldspar and commonly altered plagioclase occur. Zircon, epipode and biotite are common accessory minerals in all three units.

Padana sandstones plot in the “undissected arc” provenance field of Dickinson (1985) (Fig. 7a), whereas Quxia sandstones plot in the “transitional arc” to “dissected arc” fields (Fig. 7a), indicating upward-increasing contributions from intrusive rocks (Fig. 7a). Jialazi sandstones plot in the “dissected arc” and “recycled orogen” provenance fields (Fig. 7a), reflecting further upward increase in quartz and mixed magmatic arc and orogenic provenance.
4.3. Geochemistry of detrital Cr-spinel

Cr-spinels were found throughout Upper Cretaceous and Paleogene strata in the Cuojiangding area (Appendix Table 3). In the Padana Formation they have Mg# [Mg/(Mg+Al)] value between 0.29 and 0.61, with corresponding Cr# [Cr/(Cr+Al)] varying from 0.49 to 0.71 (0.87 in one grain). TiO$_2$ content is $\leq$ 2.45%. Detrital Cr-spinels in the Quxia Formation are indistinguishable, with Mg# between 0.26 and 0.62, Cr# between 0.46 and 0.70, and TiO$_2$ $\leq$ 2.03%. Cr-spinels in the Jialazi Formation show a wider range of Cr# (0.45 – 0.96), with Mg# mostly between 0.24 and 0.69 (as low as 0.1 for some grains), and TiO$_2$ $\leq$ 2.75%.

Cr-spinels from different rocks and different tectonic settings can be differentiated by chemical composition and binary plots of TiO$_2$ versus Al$_2$O$_3$ or TiO$_2$ versus Cr# [Cr/(Cr + Al)] (e.g. Dick and Bullen, 1984; Arai, 1992). Most spinels from residual mantle peridotites tend to have lower TiO$_2$ (<0.2%) than spinels from volcanic rocks (Kamenetsky et al., 2001). Overall, 72% detrital Cr-spinels from the Quxia and Jialazi formations have <0.2% of TiO$_2$, suggesting a mantle-peridotite source, whereas 11% Cr-spinels have >1.0% TiO$_2$ and plot into the “intraplate or oceanic island basalts” field (Dick and Bullen, 1984) (Fig. 8), suggesting a basaltic source. It is noteworthy that Cr-spinels from the Paleogene Quxia and Jialazi formations are similar to those in Cretaceous Xigaze forearc strata, but distinct from those in Yarlung Zangbo ophiolites (see Hu et al., 2014 and references therein) (Fig. 8).

4.4. Detrital zircon U–Pb ages and Hf isotopic ratios

Seven sandstone samples were collected for detrital zircon U–Pb dating (Fig. 3), as listed in Appendix Table 1. Samples 10CJD02 and 10CJD03 are from the lower Padana Formation, sample 09QXB03 from the base of the Quxia Formation, and two samples 09JLZ07 and 09JLZ08 from the top of the Jialazi Formation. Data are listed in Table 1, and age spectra and concordia diagram are shown in Fig. 9 and Appendix Fig. 2, respectively.

From the two Padana samples, 134 grains were dated, and 132 usable ages obtained. Zircons are all Cretaceous in age (104–74 Ma), excepting one Mesoproterozoic grain (1200 ± 13 Ma with discordance of 0.1%). The main age clusters are around 84 Ma and 92 Ma (Fig. 9).

From the Quxia sample, 79 zircon ages were obtained. Most zircons (94%) are Cretaceous in age (100–65.5 Ma), but 5% are Jurassic (163–146 Ma) and one Neoproterozoic (775 ± 10 Ma) (Fig. 9). The main peaks are 85 Ma and 90 Ma, with two minor peaks at 66 Ma and 75 Ma.
From the two Jialazi samples, 135 grains were analyzed and 126 usable ages obtained. Among them, 76% (n = 96) are Jurassic to Paleogene and mainly fall into three groups: 54–70 Ma (n = 14), 72–128 Ma (n = 69) and 147–186 Ma (n = 10) (Fig. 9; Table 1). Pre-Ordovician zircons are also common (23%; n = 29) and define four clusters: 503–555 Ma (n = 4), 877–1299 Ma (n = 14), 1427–1667 Ma (n = 6), 1817–1828 Ma (n = 2), with oldest grains of 2513 Ma and 2142 Ma (Fig. 9). The youngest ages are 54.0 ± 0.9 Ma, 54.1 ± 0.9 Ma, 54.0 ± 2.0 Ma, with discordance of 2%, 7%, 2%, respectively. Other young ages are 56.7 ± 0.9 Ma and 57 ± 1 Ma.

Zircon grains with age <250 Ma from the Quxia and Jialazi formations (n = 67) and Sangdadinlan and Zheda formations in Saga (n = 28) can be separated into two populations, characterized by moderately negative to positive (Group 1) and markedly negative εHf (t) values (Group 2) (Fig. 10; Appendix Tables 5).

Group 1 zircons are invariably dominant (91% in the Quxia and Jialazi formations, 100% in the Sangdadinlan and Zheda formations), having U–Pb ages from 54 to 201 Ma and εHf (t) from −2.5 to +17.0 in the Quxia and Jialazi formations, and from 49 to 70 Ma and εHf (t) from −8.4 to +13.0 in the Sangdadinlan and Zheda formations. Group 2 zircons (9% in the Quxia and Jialazi formations) have U–Pb ages from 92 to 185 Ma and εHf (t) from −7.6 to −19.4. Group 1 zircons with generally positive εHf (t) values indicate a predominantly depleted mantle magma source, whereas Group 2 zircons with most negative εHf (t) indicate a chemically evolved magma source, such as that produced by repeated melting of continental crust.

5. Discussion

5.1. Age constraints and provenance analysis

Biostratigraphic and geochronological dating converge to indicate conclusively that deposition of the Jialazi Formation took place very close to the Paleocene/Eocene boundary. The weighted-mean plateau whole-rock 40Ar/39Ar ages of 62.6 ± 0.6 Ma and 62.0 ± 1.0 Ma from the two tuff layers in the lower Jialazi Formation and a weighted-mean zircon U–Pb age of 62.0 ± 2.2 Ma from one tuff sample (n = 11; Ding et al., 2005) are only apparently in contrast with the youngest age peak of 54.9 ± 0.7 Ma that we have obtained, because the weighted age in Ding et al. (2005) was obtained by considering all zircon-grain ages. If only the youngest peak is considered, then the weighted age recalculated from their data becomes 56.3 ± 2.1 Ma (n = 4) (Fig. 6). The youngest detrital zircon U–Pb ages in the uppermost Jialazi Formation are 54 ± 1 Ma, and YDZ, YSG and YC1 εYor ages are all very close to 54 Ma (Table 1). The maximum sedimentation age as constrained by youngest detrital zircons is thus consistent with the sedimentation age indicated by foraminiferal fauna in the upper Jialazi Formation (SBZ5, 56–55 Ma; see Section 4.1.4. above). This indicates continuous magmatic activity in the adjacent continental arc, and confirms detrital-zircon geochronology as a most effective stratigraphic tool in forearc–basin successions (Dickinson and Gehrels, 2009).

Because age-diagnostic fossils are lacking, we cannot date the age of the Quxia Formation directly. However, youngest detrital-zircon U–Pb ages indicate a maximum deposition age of 66 Ma (Maastrichtian–Danian boundary) for the Quxia Formation, which is coherent with the late Maastrichtian age of the underlying Qubeiya Formation. The minimum age of the Quxia Formation is constrained by the deposition age of the overlying lower Jialazi Formation, containing larger benthic foraminifera of late SBZ3 (58–57 Ma) age in unit 15 of the Quxia section. Considering the conformable contact between the Quxia and Jialazi Formation and typically rapid accumulation of fan-delta deposits, we conclude that both units were deposited between ~58 and 54 Ma.

Sandstones of the Cuojingding section are characterized by abundant anastatic and rhythmic detritus derived from a volcanic arc. South-directed paleocurrents indicate a northern source (Ding et al., 2005), the Lhasa block with the active Gangdese arc being the most obvious candidate. This is supported by detrital zircon U–Pb spectra (Fig. 9), which compare well not only with those of Xigaze forearc-basin sandstones (Wu et al., 2010; Aitchison et al., 2011) but also with the compilation of most reliable zircon ages obtained from 219 samples of intrusive and volcanic rocks in the Lhasa terrane (Wen et al., 2008; Ji et al., 2009; Zhu et al., 2011; database summarized in Hu et al., 2012). The distinct Hf isotopic signature of Group 1 and Group 2 detrital zircons from the Quxia and Jialazi formations (Fig. 10) indicates provenance not only from the southern Lhasa subterrane (Gangdese arc), characterized by 215 to 39 Ma zircons with εHf (t) > 0, but also from central Lhasa subterrane, characterized by 216–101 Ma zircons with more negative εHf (t) from −17 to +3 (database summarized in Zhu et al., 2011; Hu et al., 2012 and references therein) (Fig. 10).

New information is provided by our geochemical data on detrital Cr-spinels in the Quxia and Jialazi formations, which are markedly distinct from Cr-spinels in the Yarlung Zangbo ophiolites (Fig. 8) and do not support provenance from the latter as inferred instead by Ding et al. (2005). Geochemical similarity of Cr-spinels in the Quxia and Jialazi formations, underlying Padana Formation, and Ngamring and Padana formations in the Xigaze area (Guo et al., 2012; An et al., 2014), rather indicates provenance from the Lhasa terrane (Hu et al., 2014). Potential sources of Cr-spinel in the Gangdese region include: 1) the Xietongmen ultramafic and mafic rock suite of debated origin, found 10 km north of the Xigaze forearc basin (Gao et al., 2003; Li et al., 2003); 2) the E/W trending and ≥60 km long Sumdo ultramafic eclogite belt of Permian age, exposed in the eastern Lhasa Block and interpreted as a dismembered ophiolite (Chen et al., 2009; Yang et al., 2009). Cr-spinels in the Quxia and Jialazi formations are unlikely to have been recycled from uplifted Cretaceous Xigaze forearc strata (Ngamring and Padana formations), because very few sandstone pebbles occur in the Quxia conglomerate, and sedimentary lithic grains (Fig. 7b) are quite rare in Quxia and Jialazi sandstones.
block, as indicated by dominant volcanic rock fragments (Fig. 7b) and ages of detrital zircons remarkably close to depositional age (99% between 104 and 74 Ma). The Quxia sandstones have less volcanic grains and consequently plot in the “transitional” to “dissected” magmatic arc provenance field (Fig. 7a). Beside zircons of Cretaceous age (100–65.5 Ma), a few Jurassic zircons appear (5% between 146 and 163 Ma), suggesting deepening of erosion into Jurassic magmatic rocks of the arc massif. Volcanic grains decrease further in the Jialazi sandstones, where quartz increases at the expense of feldspars and detrital zircons display a much wider age range (76% between 54 and 186 Ma, 23% older than Ordovician). We can thus safely conclude that the Jialazi Formation was derived not only from Jurassic to Paleogene arc rocks, but also from much older siliciclastic rocks of the Lhasa terrane, where Pre-Ordovician zircons occur in abundance (Leier et al., 2007b). Our data thus robustly document progressive unroofing of the Gangdese arc and Lhasa block, from undissected stage in the latest Cretaceous (Padana Formation), to partially dissected stage in the Paleocene (Quxia formations), and final transition to dissected stage close to the Paleocene/Eocene boundary (Jialazi Formation).

5.3. Paleogene regional correlation

The Sangdanlin succession of the northern Tethyan Himalaya (exposed in fault contact with the surrounding mélangé near Saga,
–25 km south of the Yarlung Zangbo ophiolites; Fig. 1a) includes quartzarenites sourced from India to the south (Denggang Formation; Fig. 11), overlain by 40–50 m-thick red cherty mudrocks of Paleocene age (Ding, 2003; Wang et al., 2011; DeCelles et al., 2014; Wu et al., 2014). Above, the 125-m-thick Paleocene Sangdanlin Formation (Ding, 2003; DeCelles et al., 2014) includes quartzo-lithic volcaniclastic turbidites containing chert, shale/siltstone and phyllite grains, overlain by similar quartzo-lithic volcaniclastic turbidites interbedded with gray mudrocks and cherts (Zheya Formation; Wang et al., 2011). U–Pb age spectra displayed by detrital zircons in both Sangdanlin and Zheya formations indicate provenance from the Lhasa terrane (Wang et al., 2011; DeCelles et al., 2014; Wu et al., 2014).

Further south, in the Tingri area of the southern Tethyan Himalaya (~70 km south of the Yarlung Zangbo ophiolitic suture (Fig. 1a), the lower Eocene terrigenous Enba Formation (Wang et al., 2002) overlies shallow-water limestones of the Zongpu Formation with slightly angular unconformity (Hu et al., 2012) and is largely derived from Gangdese arc rocks of the Asian margin (Wang et al., 2002; Zhu et al., 2005; Najman et al., 2010). The age of the Enba Formation is constrained by calcareous nannofossils and planktonic foraminifera as early Eocene (53.5–50.4 Ma; zones NP11–12 and P7–8; Zhu et al., 2005; Najman et al., 2010).

Very similar framework petrography (Fig. 7c, d), Cr-spinel geochemistry (Fig. 8), U–Pb age spectra and Hf isotopic ratios of detrital zircons (Figs. 9 and 10) in the Quxia and Jialazi s, Sangdanlin and Zheya formations in Saga (Wang et al., 2011; DeCelles et al., 2014; Wu et al., 2014), and Enba and Zhaguo formations in Tingri (Hu et al., 2012) suggest deposition in the same sedimentary basin and largely with the same sediment sources. Detritus was most probably derived from the same source area as for the western Xigaze forearc basin, whereas a distinct provenance characterized the eastern Xigaze forearc basin, as indicated by the detrital zircon population diagram in Fig. 9.

5.4. Paleotectonic significance of the Quxia and Jialazi formations

The Paleogene Quxia and Jialazi formations, overlying the uppermost Cretaceous Padana and Qubeiya formations (Fig. 12a), may be interpreted either as the continuation of the Xigaze forearc succession (Model 1; Fig. 12b), or as deposited in a syncollisional basin developed during the earliest stages of the Himalayan Orogeny (Model 2; Fig. 12c).

5.4.1. Paleogeographic model 1

Several authors have suggested that the Quxia and Jialazi formations represent continuing sedimentation in the Xigaze forearc basin (Liu et al., 1988; XBGMR, 1997; Jia et al., 2005; Wang et al., 1999, 2012; Fig. 12b). This implies that the India–Asia collision took place after deposition of the Jialazi Formation (i.e., after 54 Ma), as favored by the paleomagnetic study in the same area by Meng et al. (2012). According to this interpretation, the supposedly angular unconformity between the Qubeiya and Quxia formations and the sedimentological changes within the Quxia and Jialazi formations must have resulted from changes in the dynamics of Neo-Tethyan oceanic subduction, as other unconformities documented in forearc basins worldwide (e.g., Neogene Cascadia forearc in western North America; McNeill et al., 2000; Cretaceous Yezo forearc in Japan; Ando and Tomosugi, 2005). The Qubeiya/Quxia unconformity, corresponding to a time gap potentially including much of Paleocene time (66–58 Ma), may be correlated with the angular unconformity between strongly deformed Mesozoic strata and overlying Linzizong volcanic succession in south...
Tibet (Burg et al., 1983; He et al., 2007; Mo et al., 2008). This would indicate that the entire Asian arc-trench system, from the Gangdese arc to the Xigaze forearc, experienced strong crustal shortening caused by an unknown tectonic event several Ma before collision with India.

5.4.2. Paleogeographic model 2

Alternatively, the Quxia and Jialazi formations may have been deposited in a syncollisional basin formed as a consequence of incipient collision between India and Asia (Fig. 12c). This is the scenario favored here, based on the following geological evidence: 1) dramatic change from shallow shelf environments (Qubeiya Formation) to debris-flow sedimentation in fan-delta settings (Quxia Formation; Liu et al., 1988; Ding et al., 2005); 2) increasing supply from Paleozoic rocks of the Lhasa continental margin as recorded by appearance of abundant zircons older than the Ordovician in the Jialazi Formation, indicating accelerated erosional unroofing of the southern Lhasa block; 3) palynofloral assemblages in the Quxia Formation (with a climax of bisaccate pollens typical of conifers living in cool mountain areas; Li et al., 2008), indicating considerable elevation of the Gangdese arc; 4) provenance similarities of the Quxia and Jialazi formations with the Sangdanlin and Zheya formations of the northern Tethys Himalaya (Wang et al., 2011; DeCelles et al., 2014; Wu et al., 2014) and with the Enba and Zhaquo formations of the southern Tethys Himalaya (G.B. Li et al., 2005; Hu et al., 2012; Q.H. Zhang et al., 2012), supporting deposition in the same syncollisional basin with the same orogenic sources (Fig. 10).

If this model is correct, then the Quxia and Jialazi formations are syncollisional deposits accumulated on top of the nascent Himalayan Orogen (Fig. 12c). The Quxia and Jialazi formations would thus represent an example of collision-related fan delta (as documented in collision orogens worldwide; e.g., Ethridge and Wescott, 1984; López-Blanco et al., 2000), supplying sediments to deep-water environments (Sangdanlin and Zheya formations; Wang et al., 2011). According to Model 2, the unconformity between the Qubeiya and Quxia formations would correspond to the very onset of the India–Asia continental collision. Collision timing would thus be constrained as younger than the Qubeiya Formation (~66 Ma) and older than the Quxia and Jialazi Formations (~58 Ma).

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**Fig. 11.** Correlation of Paleogene marine strata along a NW/SE transect in central south Tibet (Fig. 1a). PFZ: planktonic foraminiferal zone; SBZ: shallow benthic foraminiferal zone after Serra-Kiel et al. (1998); time scale after Gradstein et al. (2012).
58-54Ma  Model 2: Syn-collisional basin after the India-Asian collision

The uncertainty remains on the origin of environmental change close to the Paleocene/Eocene boundary from proximal fan-delta deposition dominated by debris-flow conglomerates to distal fan-delta shallow-marine sedimentation. This transgression may reflect either regional tectonic causes or possibly even global eustatic control associated with latest Paleocene–early Eocene greenhouse warming (Zachos et al., 2008).

6. Conclusions

Integrated stratigraphic, sedimentological, petrographic, geochronological and geochemical data on the uppermost Cretaceous to Paleogene succession exposed in the Cuojiangding area of southern Tibet allowed us to unravel the tectonic events immediately preceding and associated with the earliest evolution of the Himalayan Orogen. The deltaic Padana Formation and the overlying inner-shelf Qubeiya Formation document the filling stage of the Xigaze forearc basin at the close of the Cretaceous. A tectonic event is testified by the Quxia and overlying Jialazi formations, deposited in fan-delta settings. A latest Paleocene–earliest Eocene age (57–54 Ma) for the Jialazi Formation is robustly constrained by foraminiferal assemblages and radiometric dating of interbedded tuff layers.

Sandstone petrography, paleocurrents, U–Pb ages and Hf isotopic ratios of detrital zircons indicate provenance from the Gangdese arc throughout the Late Cretaceous to Paleogene, with rapid unroofing of its plutonic roots during deposition of the Quxia Formation. Erosion finally cut into deeper-seated older rocks of the Lhasa Block during deposition of the Jialazi Formation. Detrital Cr-spinels from the Upper Cretaceous to Paleogene succession in Cuojiangding area compare well with those contained in Xigaze forearc strata but poorly with those in Yarlung Zangbo ophiolites, confirming provenance from the Lhasa terrane.

The dramatic change from shelfal sedimentation in the latest Cretaceous (Padana and Qubeiya formations) to fan-delta sedimentation in the Paleogene/Quxia and Jialazi formations, and the close affinity in sandstone petrography, Cr-spinel geochemistry, U–Pb ages and Hf ratios of detrital zircons in the Jialazi Formation and both Sangdanlin + Zheya formations of the northern Tethys Himalaya and Enba + Zhaguo formations of the southern Tethys Himalaya testify in favor of paleogeographic Model 2. The Quxia and Jialazi formations are thus interpreted to represent syncollisional sediments deposited just
after the onset of the India–Asia collision on top of the nascent Himalayan orogenic belt.

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Appendix A. Supplementary data
Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.gr.2015.02.007.

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