Dating the India–Eurasia collision through arc magmatic records

Pierre Bouilhol a,1, Oliver Jagoutz a,* , John M. Hanchar b , Francis O. Dudas a

a Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139, USA
b Department of Earth Sciences, Memorial University of Newfoundland, St. John’s, Newfoundland, Canada A1B 3X

1 Present address: Department of Earth Sciences, Durham University, Durham DH1 3LE, United Kingdom.

* Corresponding author. Tel.: +1 617 324 5514. 
E-mail addresses: pierre.bouilhol@durham.ac.uk (P. Bouilhol), 
jagoutz@mit.edu (O. Jagoutz), jhanchar@mun.ca (J.M. Hanchar).

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ABSTRACT

The Himalayan orogeny, a result of the collision of India and Eurasia, provides direct evidence of strain accommodation and large-scale rheological behavior of the continental lithosphere. Knowledge of the timing of the India–Eurasia collision is essential to understand the physical processes involved in collisional systems. Here we present a geochronological and multi-isotopic study on rocks from the upper crust of the Kohistan Paleo-Island Arc that formed in the equatorial part of the Neo-Tethys Ocean. In situ U–Pb geochronology and Hf isotopes in zircon, and whole-rock Nd and Sr isotopic data of plutonic rocks from the Kohistan-Ladakh Batholith, are used to construct a continuous record of the isotopic evolution of the source region of these granitoids that are related to both the subduction of the oceanic lithosphere and subsequent arc–continent collisions. We demonstrate that profound changes in the source region of these rocks correspond to collisional events. Our dataset constrains that the Kohistan–Ladakh Island Arc initially collided along the Indus suture zone with India at 50.2 ± 1.5 Ma, an age generally attributed to the final India–Eurasia collision for the entire Himalayan belt. In the western Himalaya, the final collision between the assembled India/Arc and Eurasia however, occurred ~10 Ma later at 40.4 ± 1.3 Ma along the so-called Shyok suture zone. We present evidence indicating that a similar dual collision scenario can be extended to the east and conclude that a final India/Arc–Eurasia collision at ~40 Ma integrates crucial aspects of the magmatic, tectonic, and sedimentary record of the whole Himalayan mountain belt.

1. Introduction

The continent–continent collision of India with Eurasia has significantly shaped the surface of the Earth by producing one of the largest known orogenic systems. To understand the importance of the physical processes controlling crustal deformation and the influence of this mountain building event on the climatic system (Le Pichon et al., 1992; Molnar et al., 1993; Molnar and Tapponnier, 1975; Royden et al., 2008; Tapponnier et al., 1982), an accurate knowledge of the plate convergence and the kinematic history of the collision is essential. Key boundary parameters include the timing of collision and plate reconstructions, which together constrain the absolute amount of plate convergence accommodated. Whereas modern paleomagnetic studies agree on plate reconstruction within a few hundred kilometers (e.g., van Hinsbergen et al., 2012), the timing of the final India–Eurasia collision is debated. Estimates range from ~70 to ~25 Ma (e.g., van Hinsbergen et al., 2012; Yin and Harrison, 2000) with most workers preferring a ~52–50 Ma collision age (e.g., Najman et al., 2010). Recent paleomagnetic data suggest a complex collision scenario that involved multiple continental or arc fragments and increasingly casts doubt that the ~50 Ma events constrains the final India–Eurasia collision (Aitchison et al., 2007; Khan et al., 2009; van Hinsbergen et al., 2012). This uncertainty on the collision age translates to up to ~3000 km differences in estimates on the convergence accommodated in the system, which strongly hampers our understanding of the formation of the India–Asia collisional system.

In this paper we present new results that constrain the timing of collision in the western Himalaya (Fig. 1). Our approach is derived from our understanding of currently active continent–island arc collision zones (see Section 3). In these systems, the subduction of continental lithosphere below the arc can be tracked by the differences in isotopic compositions between pre- and post-collisional arc magmas if significant isotopic differences exist between the subducting continental lithosphere and the ombducting arc crust. Along the Himalayan belt, only the Kohistan–Ladakh Paleo-Island Arc (KLA) that is exposed in the western Himalaya offers such an opportunity.

The main objective of this paper is to construct a continuous spatial and temporal record of the isotopic composition of the KLA related rocks, spanning from the intra-oceanic subduction in the early Cretaceous to the Miocene, thus covering all proposed collision times. Kohistan–Ladakh Arc magmas that formed prior
to the 50–52 Ma collision of the arc with India have isotopic characteristics comparable to modern intra-oceanic arc volcanics (Figs. 2 and 3). Conversely, the post-collisional, late-Eocene/Oligocene granites are more variable in composition and trend to significantly more radiogenic isotopic compositions (Heuberger et al., 2007) indicating the involvement of old continental lithosphere in their formation.

2. Geological setting

Three suture zones, the Indus, Shyok and Tsangpo, mark the Himalayan collision (Fig. 1). In the NW Himalaya, the ~70,000 km² KLA is wedged between India and Karakoram (the former paleo-Eurasian continental margin). The Shyok suture separates the KLA in the north from the Karakoram, whereas in the south the Indus suture divides the KLA from the Indian continent (Fig. 1). The Tsangpo (or Yarlung-Tsangpo) separates India from Eurasia in the central and eastern part of the Himalaya (Burg and Chen, 1984; Gansser, 1980; Yin and Harrison, 2000).

2.1. The KLA and the Karakoram

The KLA (for a recent review see Burg, 2011) formed primarily during the late Mesozoic as an intra-oceanic arc, during the northward directed intra-oceanic subduction of the Neotethys ocean basin (Bard, 1983; Coward et al., 1987; Tahirkheli, 1979). This interpretation is based on the absence of continental basement and continental derived sediments in the KLA (Bard, 1983; Jagouzt et al., 2009; Tahirkheli, 1979), and on paleomagnetic data placing the formation of the KLA near the equator within an oceanic basin (e.g., Khan et al., 2009) much further south than the Eurasian margin during the Mesozoic (Schettino and Scotese, 2005).

To the north of the arc system, at the northernmost margin of the Neotethys, a second subduction system was active simultaneously, resulting in the Jurassic to Tertiary Gangdese and Karakoram Batholiths (e.g., Debon et al., 1987; Gaetani, 1997; Heuberger et al., 2007; Searle et al., 2010). These Batholiths were part of an active continental margin due to the northward subduction of Tethyan oceanic lithosphere beneath the southern Eurasian margin (e.g., Le Fort, 1988; Rex et al., 1988; Yin and Harrison, 2000).

2.2. The Indus suture

In contrast to the Tsangpo suture, the Indus suture zone contains eclogites and ultra-high pressure (UHP) assemblages that can be used to reliably constrain the timing of the formation of the Indus suture. The P-T-t path of the eclogites provides evidence for the beginning of the subduction of the Indian margin at ~50 Ma (Kaneko et al., 2003; O’Brien et al., 2001; Parrish et al., 2006). In conjunction with the fact that the sedimentary record placed the transition from marine to non-marine sedimentation on the northern Indian Margin at ~50 Ma (e.g., Beck et al., 1995; Garzanti et al., 1987; Green et al., 2008), the timing of the Indus suture, and the collision between India and the KLA, have been accepted to be ~50 Ma.

2.3. The Shyok suture

The Shyok suture has a complex history and its formation is poorly constrained. North verging folds and top-to-the-north directed shear zones within the northern part of the KLA indicate that the KLA was initially overthrust on top of the Karakoram margin (Coward et al., 1986). During the early Miocene, parts of the Shyok suture were reactivated as a top-to-the-south directed thrust, overprinting the previous suture related structures (Brookfield and Reynolds, 1990; Coward et al., 1986).

Originally, the formation age of the Shyok suture was interpreted to postdate the India–KLA collision (Achache et al., 1984; Andrews speed and Brookfield, 1982; Bard, 1983; Brookfield and Reynolds, 1981). Subsequently, based on a proposed systematic relationship between emplacement age of plutonic rocks from the northern KLA (determined by Rb–Sr whole rock and mineral isochrons) and the presence or absence of deformational fabric in the rocks (i.e., younger ones being undeformed, older deformed), the Shyok suture has been inferred to have formed during the
Cretaceous (Petterson and Windley, 1985; Windley, 1988). However, recent U–Pb zircon dates demonstrate that no systematic relationship between intrusion ages and absence or presence of deformational fabric exists (Jagoutz et al., 2009), leaving the formation of the Shyok suture zone essentially unconstrained (Burg, 2011; Jagoutz et al., 2009). Similarly, paleomagnetic data indicate that the KLA was situated ~3000 km south of the Eurasian margin in the Tertiary, making a Cretaceous formation age of the Shyok suture highly unlikely (Ahmad et al., 2000; Khan et al., 2009).

3. Approach for dating the India–KLA–Eurasia collision events

The rationale of our method to constrain collision events of the KLA by investigating the isotopic composition of igneous rocks follows from observations on modern continent–arc collision systems and the documented impact of collision on the geochemistry of the arc-related magmas. The Australian–Sunda–Banda Arc collision system, for example (see detailed review by Harris, 2011) provides a well-documented modern analog for the India–KLA collision. The Australian continent, drifting north–northeastward (NNE) at a rate of ~70 mm/a (Nugroho et al., 2009), initially collided...
<table>
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<th>Rock type</th>
<th>Age (Ma)</th>
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| Northern section | | | | | | | |
| PK05-39 | 3983738 | Grano. | 86.0 | PK04-16 | 4006683 | Granodiorite | 83.5 |
| PGC-16 | 342237 | Grano. | 83.5 | PK04-16 | 4006683 | Granodiorite | 83.5 |
| PGLT-19 | 398516 | Quarts-diorite | 64.8 | PK06-38 | 4006708 | 6384781 | Monzogranite | 51.5 |
| PK04-9 | 300736 | Grano. | 58.9 | PK09-16.6A | 3708986 | 7696865 | Monzogranite | 60.9 |
| PK09-16.6A | 3708986 | Grano. | 59.5 | PK09-11.4A | 3975084 | 792703 | Granodiorite | 39.8 |
| PK09-11.5A | 3716910 | Grano. | 30.1 | PK06-36 | 3892534 | Kfs-Granodiorite | 41.5 |

- Error number corresponds to the last digit.
- Refers to samples that have a heterogenous Zircon population.
- Values in Italic refer to isotopic dilution TIMS (ID-TIMS) measurements.
changes. As shown below, simple mixing calculations of partial variations in the sources and the significance of any observed isotope ratios (e.g., $^{143}\text{Nd}/^{144}\text{Nd} = 0.51278$) in the formation of the granitoids, we modeled their isotopic compositions reflecting the formation of the melts in a slab-contaminated mantle wedge similar to the current situation further west in Java (Handley et al., 2007) and typical for intra-oceanic arcs (e.g., Elburg et al., 2005). In the back-arc side of the collision zone, the furthest from the trench, the isotopic composition of contemporaneous magmas show little variation ($^{143}\text{Nd}/^{144}\text{Nd} \sim 0.5125(1)$), because the Australian continent has not been subducted deep enough to influence the source region of these melts (Elburg et al., 2005). Accordingly, the subduction of the Australian lithosphere can be closely tracked in space and time by the chemical composition of the related magmatic rocks. Based on these observations we have investigated the possible geochemical effects of the India–KLA-Eurasia collision on the arc-related magmatic products.

Within the Himalayan system, only the KLA offers this opportunity as it formed as a Cretaceous to Eocene intra-oceanic island arc and accordingly lacks the continental basement of the Karakoram or Lhasa Terranes (e.g., Chung et al., 2009; Yin and Harrison, 2000 and references therein) allowing attribution of isotopic changes to the deeper source region of the rocks studied, rather than, for example, changes in intracrustal assimilation processes.

4. Analytical methods

We used in situ zircon U–Pb geochronology and Hf isotopes (Table A.1; Fig. A.1), and whole-rock Nd and Sr tracer isotopes of 41 granitoids samples from the KLA batholiths ranging in composition from diorite to granite (Table 1). For each sample, U–Pb dates and Lu–Hf compositions were determined at the Inco Innovation Centre of the Memorial University of Newfoundland (St. John’s, Canada). U–Pb and Pb isotopic ratios were analyzed using a Finnigan ELEMENT XR ICP-MS coupled to GeoLas 193 nm ArF Excimer laser. A 10 μm laser beam was rastered over the zircon surface in a 40 × 40 μm grid. The in situ Lu–Hf isotopic analyses were acquired using a Finnigan NEPTUNE multi-collector ICP-MS directly on top of the U–Pb raster using a 49 or 59 μm laser spot using the same laser ablation system. Bulk-rock isotopic analyses were done at MIT using standard wet chemical separation techniques and the IsoProbe-T multicollector thermal ionization mass spectrometer (TIMS). See Electronic supplement for more detailed information about the analytical methods used in this study.

To identify and constrain the possible end-members involved in the formation of the granitoids, we modeled their isotopic compositions (Table 2; Fig. 4). This was done in order to identify variations in the sources and the significance of any observed changes. As shown below, simple mixing calculations of partial melts (calculated as batch melts) derived from four different source rocks, successfully explain the observed variations (Table 2; Fig. 4): (1) Enriched–Depleted MORB Mantle (E–DMM); (2) Tethys-type MORB; (3) evolved continental crust; and (4) the KLA lower crust. This fourth potential source rock does not represent a different independent isotopic end-member component, as its composition lies on a mixing line between Tethyan MORB and E–DMM. We chose E–DMM for the mantle component because of the enriched isotopic character of the mantle in the Indian Ocean, which is reflected in the isotopic composition of the arc (Bouilhol et al., 2011; Jagoutz and Schmidt, 2012; Khan et al., 1998). On average, the granitoids composition are close to a melt composition and cumulates are rare (Jagoutz et al., 2011), thus allowing to compare directly their measured Sr and Nd concentrations in our model. To model the Hf and Lu data we used mineral/melt Kd data (1000 °C zircon Kd from Rubatto and Herrmann, 2007) to calculate the Lu and Hf concentrations of the corresponding liquids in equilibrium with the analyzed zircon grains. With the exception of a depleted mantle component (E–DMM), all other source rocks have been constrained from actual geochemical data (e.g., trace elements and radiogenic isotopes) from rocks exposed in the KLA and surrounding units or from relevant geographical areas, providing very specific constraints for our model calculations. All parameters used for the model calculations are provided in Table 2.

The spatial distribution of the samples investigated covers the entire KLA batholith and define three sections (Fig. 1): (1) the southern section comprises the samples that are the closest to the Indus suture; (2) the northern section consists of the samples closest to the Shyok suture; and (3) the central section include all other samples that are geographically in between the two previously defined sections. For each section, the sample’s isotopic record ($^{143}\text{Nd}/^{144}\text{Nd}, ^{87}\text{Sr}/^{86}\text{Sr}$, and $^{176}\text{Hf}/^{177}\text{Hf}$) are discussed in relation to their U–Pb zircon crystallization age (Figs. 2, 3 and 5).

5. Results and discussion

In regard to their isotopic compositions, the KLA granitoids define two broad groups: (1) an older depleted group that reflects a juvenile character of the granitoid source; and (2) a younger more isotopically evolved group indicating the contribution from continental crust. As the transition between the different groups occurs at different times and at different places in the arc, we discuss the results separately for the southern, northern, and central part of the Batholith.

5.1. The southern section

Along the Indus suture, the granitoids show contrasting isotopic records that can be temporally divided into two suites of samples: samples that are older than ~50 Ma, show an isotopic record that is typical of the intra-oceanic history, whereas those that are younger than ~50 Ma present a wide range of composition that involves an old continental component in their source. The results of these two suites of samples are thus shown and discussed separately.

5.1.1. Pre-50 Ma samples

From 102.1 ± 2.1 to 50.3 ± 2.1 Ma, each granitoid sample from the southern part has a homogeneous U–Pb zircon age population that constrains the crystallization age (Fig. 2). Some samples contain rare inherited zircon grains ranging in age from 94 to 62 Ma (Table A.1). The inter-grain zircon $^{176}\text{Hf}/^{177}\text{Hf}$ variations in a given sample, overlap within analytical uncertainty, allowing us
to calculate an average \( e^{Hf} \) for each sample (Fig. 2; Table A.1). The \( e^{Hf} \) variability between samples shows a restricted range of \( 6.2 \pm 1 \) to 12.1 ± 1.1 (Fig. 2) with an average of 9.4 ± 0.7. Inherited older zircon grains exhibit a range in \( e^{Hf} \) comparable to the variability between samples and are attributed to reworking and assimilation of older arc-crust. The whole rock \( ^{187}Nd/^{144}Nd \) compositions of the same samples show an \( e^{Nd} \) that ranges from 1.2 to 5, with a weighted mean of 2.6 ± 0.7 (Table 1; Figs. 2 and 3), and an initial \( ^{87}Sr/^{86}Sr \) ranging from 0.703744 to 0.704719. The \( e^{Hf} \), \( e^{Nd} \) and \( ^{87}Sr/^{86}Sr \) observed during this time interval is within the range of the isotopic composition previously reported for the KLA rocks of this age.

The modeling results of the isotopic and trace element characteristics of these rocks indicate that the observed variations are...
samples have heterogeneous zircon populations dominated by inherited grains (Fig. 2). The youngest grains that were used to calculate the crystallization age often show significant inter-grain $^{176}$Hf/$^{177}$Hf variations of up to 10 $^{144}$Nd/$^{143}$Nd, as well as an age of −15. The $^{144}$Nd/$^{143}$Nd for <50 Ma samples is negative and varies over 6 $^{144}$Nd/$^{143}$Nd (−9.7 < $^{144}$Nd/$^{143}$Nd < −3.6), associated with very radiogenic Sr isotopic composition of $^{87/86}$Sr, ranging from 0.705862 to 0.713170 (Figs. 2 and 3). The youngest heterogeneous sample (LB09-11.5A, 29.6 ± 0.8 Ma), having the most isotopically evolved composition, contains numerous inherited zircon grains that yield generally discordant Paleozoic $^{206}$Pb/$^{238}$U dates between ~434 Ma and ~384 Ma and $^{207}$Pb/$^{206}$Pb dates between ~940 and ~474 Ma (Table A.1). These older grains have maximum Paleozoic/Proterozoic $^{144}$Nd/$^{143}$Nd values between −7 and −2, comparable to that of the young zircon grains. The age and isotopic composition of the inherited zircon grains are similar to those found in the Indian margin (Veevers and Saeed, 2009), but are absent from the >50 Ma KLA rocks. A significant shift of ~10 $^{144}$Nd/$^{143}$Nd and ~6 $^{144}$Nd/$^{143}$Nd units between the more unradiogenic and the isotopically evolved samples is observed at 50.2 ± 1.5 Ma, together with a significant increase in Sr isotopic composition. This distinct change is coeval within error with the onset of the India–KLA collision (Garzanti et al., 1987; Kaneko et al., 2003).

In contrast to the >~50 Ma samples, these granitoids display a strong variability in their isotopic composition for their limited variations in the parent and daughter elements concentrations (Fig. 4). Modeling of the trace element and isotopic compositions of these isotopically evolved rocks does not suggest direct involvement of a depleted mantle component in their formation (Fig. 4), but rather require the involvement of an additional enriched component, such as old isotopically evolved crust. Additionally, the isotopically evolved composition necessitates the involvement of an arc component such as the lower KLA crust (Fig. 4). The isotopic compositions of these <50 Ma rocks from the southern transect are best explained as mixtures of melts of Indian-derived sediments or the Indian crust and the lower KLA crust (Fig. 4).

The temporal concurrence of the change in isotopic compositions in the granitoids from the southern KLA and the onset of the India–KLA collision, combined with the inherited zircon grains and the isotopic record of the evolved rocks, identify the subducting Indian margin as contributing to the source of the <~50 Ma rocks. Therefore, the shift in isotopic compositions at 50.2 ± 1.5 Ma reflects a profound change of the source regions of the KLA magmas associated with the India–KLA collision. The expected delay between the initial contact of India with the KLA and the formation of evolved granites is unresolvable within our analytical uncertainty and thus less than ~1.5 Ma, corresponding to ~150–200 km of subducted Indian lithosphere.

5.2. The northern section

In contrast to the southern part, samples from the northern section maintain a more depleted isotopic signature until ~40 Ma (Fig. 2). Samples that are >40 Ma show an isotopic record that defines their juvenile character, whereas those that are <40 Ma have isotopic characteristics that require the presence of an old continental component in their source. Therefore, like for the southern section, the granitoids samples from the northern section are presented and discussed in two steps.

5.2.1. Pre-40 Ma samples

From 86 ± 2.6 Ma to 40.9 ± 1.1 Ma, these samples have homogeneous zircon U–Pb age populations resulting in tightly

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**Fig. 5.** For the Central Batholith samples, U–Pb zircon crystallization ages of granitoids vs. (1) their initial epsilon Hafnium ($^{144}$Nd/$^{143}$Nd) composition of zircon grains and (2) initial epsilon neodymium isotopic whole rock composition ($^{144}$Nd/$^{143}$Nd) during the intra-oceanic history of the arc. They show large island arcs, indicating their juvenile character and their formation from the E–DMM and from subducted oceanic lithosphere (Fig. 4). The involvement of these components in their source, typical for island arcs, indicates their juvenile character and their formation during the intra-oceanic history of the arc. They show large variability in the parent and daughter trace elements (i.e., whole rock Sm, Nd, Sm/Nd, Rb, Sr, Rb/Sr, and zircon Lu/Hf ratios) at relatively constant depleted isotopic compositions (i.e., high $^{144}$Nd/$^{143}$Nd and $^{176}$Hf/$^{177}$Hf, and low $^{87}$Sr/$^{86}$Sr(i)) that could only be partially explained by magmatic fractionation (Fig. 4), if they were derived from a similar parental melt composition (Fig. 4). Additionally, the major element chemistry of most of these granitoids is relatively similar, indicating only limited variations related to fractionation. Variable degrees of partial melting of one source component of the granitoids cannot explain the observed variations in parent and daughter element concentrations. The scatter in the isotopic composition of the data orthogonal to this mixing line (Fig. 4) could indicate a minimal contribution (<5%) of an evolved continental crust component, but can also be adequately explained by the observed variations in E–DMM and MORB end-members (Mahoney et al., 1998; Workman and Hart, 2005; Fig. 4). Therefore, the variations observed are best explained by variations in the basaltic parental magma composition, which reflects changes in the mantle source region of the basalts due to mixing of different components.

5.1.2. Post-50 Ma samples

In contrast, samples from the southern part that are younger than 50.4 ± 1.6 Ma have more evolved isotopic compositions compared to the > ~50 Ma samples. Three of the four younger
constrained U–Pb zircon crystallization ages for each sample. Rare inherited zircon grains range in age between 108 and 47 Ma. The zircon εHf(t) of the samples studied does not vary significantly and has a weighted mean εHf(t) of 10.7 ± 1.4 (Fig. 2), similar to the range observed in the inherited zircon grains. The bulk-rock εNd(t) of these samples varies between 3.5 and 5.7, and show a 87Sr/86Sr(t) between 0.703582 and 0.704344 (Figs. 2 and 3), similar to the pre-50 Ma samples of the southern section. Isotopic modeling confirms juvenile depleted mantle and basaltic oceanic crust components in the source region (Fig. 4), similarly to the > 50 Ma rocks of the southern section.

5.2.2. Post-40 Ma samples

Contrastingly to the > 40Ma samples, zircon grains from granitoids that are 39.9 ± 1.4 and 37.2 ± 1.1 Ma show a wider εHf(t) range, from -5.2 ± 1.0 to 10.1 ± 2.3, and a slightly more evolved whole-rock εNd(t) (e.g., 1.5 and 1.7) and 87Sr/86Sr(t) (0.705017 and 0.704526; Figs. 2 and 3). These granitoids contain numerous inherited zircon grains of Mesozoic age with generally concordant U–Pb dates ranging from ~235 to ~103 Ma, and εHf(t) ranging from -6.4 ± 1.4 to 1.4 ± 1.4. The ages and isotopic composition of the Mesozoic inherited zircon grains are unknown for > 40 Ma rocks from the KLA and from the northern margin of India (Veever and Saeed, 2009).

Although more evolved isotopically than the > 40 Ma rocks, these samples are significantly less radiogenic than the < 50 Ma rocks from the southern section (Fig. 2). It is noteworthy, that unlike the > 40 Ma rocks, the observed compositions do not indicate direct involvement of an E–DMM mantle component in their formation (Fig. 4). Importantly, the inherited Mesozoic zircon grains are identical in their Lu/Hf and εHf(t) isotopic compositions to zircon grains from Mesozoic plutonic rocks from the Karakoram (Figs. 2 and 4), suggesting that the Karakoram margin contributed to the source of these rocks. Indeed, the Karakoram block, of continental origin, has a multi-phase history involving Andean-type and rift related magmatism (Debon et al., 1987; Rolland et al., 2000, 2000a). During these different geodynamic episodes, significant volumes of mantle-derived melts were emplaced in the Karakoram block. Compared to the Gondwana derived basement (e.g., Crawford and Searle, 1992; Yin and Harrison, 2000) into which they intruded, this magmatism has an unradiogenic composition, thereby changing the bulk isotopic composition of the Karakoram Terrane as a whole.

The several episodes of Andean-type magmatism and rifting have generated rocks with a wide range of isotopic compositions within the Karakoram block. The presence of rocks with similar non-radiogenic Nd and Sr isotopic compositions as the < 40 Ma granitoids within the Karakoram indicate that they may have contributed to their source. In Fig. 4 we have presented the data that exist for the Karakoram crust to demonstrate its composite and complex nature. Due to the lack of data for what would have been the Karakoram margin prior to collision, and considering the wide range of compositions found in the Karakoram, it remains difficult to choose a unique end-member for the modeling of the < 40 Ma granitoids from the North. As no whole rock geochemical data from the Karakoram crust exists in the literature from which both Lu and HF concentrations and HF isotopic data are presented, we are not able to reliably model the Lu–HF system. However, we consider the inherited zircon record of these granitoids, which are identical in U–Pb age, HF isotopic composition and Lu–HF ratio to those recorded in Karakoram granitoids, as the strongest argument for the involvement of the Karakoram margin in the source of isotopically evolved rocks in the North (Fig. 4d).

Therefore, the isotopic shifts occurring at 40.4 ± 1.3 Ma along the Shyok suture is best explained by the involvement of Mesozoic rocks from the Karakoram and not the northern Indian margin. This interpretation is further supported by regional studies that have documented that the Karakoram margin was initially underthrust below the KLA (e.g., Coward et al., 1986).

Along the Indus suture, and observed in modern collisional systems (Section 3), an abrupt change from an unradiogenic to a more evolved isotopic signal in magmatic rocks closely corresponds to the onset of collision and suturing in continent–arc collision systems. We conclude that the abrupt shift in isotopic composition of approximately 16 εHf(t) units and 4 εNd(t) units observed in granitoids in the northern KLA correlates within error to the formation age of the Shyok suture, and dates the final collision between the now assembled India/KLA with the Karakoram at 40.4 ± 1.3 Ma. This final collision age implies that a ≥ 1000 km wide oceanic basin previously situated between the KLA system and the Karakoram margin was subducted after the initial India-Arc assembly, which is in agreement with the paleomagnetic data (Khan et al., 2009) and in accordance with the magmatic and metamorphic history of the Karakoram (Searle et al., 2010).

Finally, the youngest granitoids, dated at 25.1 ± 1.3 and 21.5 ± 1.2 Ma show the most isotopically evolved composition in the Northern section. They have a heterogeneous zircon population, showing a wide range of εHf(t) between −5.2 and 1.2 (Fig. 2), together with inherited grains that are Precambrian and Mesozoic in age resembling those found in the Karakoram block. The bulk-rock εNd(t) data for these samples are clustered around –6, and they show an evolved 87Sr/86Sr(t) composition (0.707422 and 0.709701; Figs. 2 and 3). These two samples have similar isotopic compositions as the Miocene ages granites from the Karakoram (Maheo et al., 2009). This Miocene post-collisional magmatism is not only present in the Karakoram, but also throughout the Tibetan Plateau, and has been widely attributed to the reworking of the Eurasian crust by the influx of hot mantle material, possibly related to a loss of crustal rock at depth (e.g., Gao et al., 2010).

5.3. The central section

Samples dated between 72.1 ± 1.4 Ma and 58.8 ± 1.4 Ma have average εHf(t) of 9.2 ± 0.6 (Fig. 5). They have a whole rock εNd(t) and 87Sr/86Sr(t) falling within the same isotopic range (Fig. 5) of the other transect for similar ages, therefore referring to the intra-oceanic history of the arc (1.7 < εNd(t) < 3.5; 0.703781 < 87Sr/86Sr(t) < 0.704507). Three samples dated between 46.3 ± 1.3 and 41.5 ± 1 Ma are intermediate in isotopic composition, showing slightly lower εHf(t) (3.9 ± 0.5 < εHf(t) < 6.8 ± 0.7) and εNd(t) (1.11) and more evolved 87Sr/86Sr(t) (0.705061) than the > 58 Ma samples. These samples do not show the pronounced isotopic shifts observed along the southern transect. A clear appearance of an old crustal evolved component in the source is only demonstrated by the two youngest granites, dated at 31.7 ± 0.8 and 30.1 ± 1.3 Ma showing a wide range of εHf(t) from 14.1 ± 1 to 1.85 ± 1.3, εNd(t) ranging from −2.50 to 1.05 and 87Sr/86Sr(t) from 0.705952 to 0.704959, after the collision of the India–Arc assembly with Karakoram (Figs. 4 and 5).

For the intermediate rock composition observed between 50 ± 40 Ma, we favor the explanation that the Indian lithosphere was initially steeply subducted into the mantle, and that the mantle wedge was still present below the central part of the arc system as shown by the isotopic composition of sample PK06-36 that involves a mantle component (Fig. 4). This interpretation is supported by the P-T-t path recorded in the ultrahigh pressure metamorphism in the western Himalaya (Guillot et al., 2008), indicating that the Indian continent was initially subducted to > 150 km. These eclogites started to be subducted at ~50 Ma, reaching their metamorphic
peak at 46 Ma, and were exhumed at 40 Ma (Kaneko et al., 2003; Parrish et al., 2006; Tonarini et al., 1993; Treloar et al., 2003). These observations confirm that the subduction of buoyant crustal material, rather than the continent-continent collision, reduced the convergence rate after 50 Ma (Capitanio et al., 2010).

5.3.1. Significance of source derived zircons
The first Indian derived zircon grains occur at ~30 Ma in the southern section, contemporaneously with the crustal component in the central section. As zircon crystals would most likely not survive in a crustal-derived melt that re-equilibrated with the mantle wedge (Hanchar and Watson, 2003), the first appearance of Indian derived zircon grains indicate the absence of a significant portion of the mantle wedge. This would be facilitated if the Tethyan oceanic slab was detached, allowing the Indian lithosphere to be underthrusted. In this model, the appearance of Indian derived zircon grains at ~30 Ma constrains the timing of the underthrusting of the Indian Lithosphere underneath the KLA. This occurred ~20 Ma later than the decrease in the convergence rate and the beginning of the Indian Lithospheric subduction, and is in accordance with numerical models (van Hunen and Allen, 2011). Accordingly, for the Northern section, underthrusting, rather than subduction, of the Karakoram margin would explain why Karakoram derived zircon grains are immediately found at the time of collision, in contrast with the southern section.

5.4. Results in the framework of the Himalayan belt
Collectively, our data, are in good agreement with paleomagnetic studies (Khan et al., 2009), and document that the KLA formed as an intra-oceanic island arc south of the Eurasian margin until its amalgamation to India at 50 Ma. This India–Arc collision has been followed by the closure of the Shyok back-arc basin at 40 Ma, corresponding to the final India/Arc–Karakoram collision in the northwestern Himalaya (Fig. 6). In the framework of the broader evolution of the India–Eurasia collision, our results can be interpreted in two main geodynamic scenarios (Fig. 6): (1) the KLA was part of a lateral extensive chain of island arcs situated in the equatorial part of the Tethys and correlate to the supra-subduction zone ophiolites associated with the Tsangpo suture zone (Aitchison et al., 2000; Hébert et al., 2012). In this scenario a second subduction zone was active to the north of the intra-oceanic system, along which the Shyok Sea was subducting underneath the Eurasian margin producing the Karakoram and Gangdese continental arcs. In this model, the India–Arc collision occurred thorough the length of this chain at ~50 Ma, and the final India/Arc–Eurasia collision occurred only at ~40 Ma; alternatively, (2), the KLA and the Gangdese formed along the same subduction zone (Yin and Harrison, 2000), similar to the modern Alaska–Aleutian subduction system. In this scenario the Shyok “Basin” formed as a marginal basin behind the KLA, and terminated toward the east, but could have extended further toward the west (Fig. 6). In this case, to explain the Karakoram continental magmatism, a second subduction zone north of the KLA–Gangdese system must have existed in the west, where the Shyok Basin was subducted underneath the Karakoram margin that has no continuation toward the east. In this model the India–Arc collision recorded in NW Himalaya at 50 Ma would correspond to the main collision to the East, and the 40 Ma reflects the closure of the marginal Shyok Basin. The subsequent discussion focuses on three main observations that have been used to constrain collision and might help to decipher between the two different scenarios.

5.4.1. Origin of the Tsangpo suture ophiolites
A fundamental differences between these two scenarios concerns the origin of the frequent ophiolites along the Tsangpo suture zone that formed in a supra-subduction zone setting (e.g., Dai et al., 2011;
The few reliable zircon U–Pb ages ranging from upper Jurassic to upper Cretaceous and Nd–Sm isochrons indicate that the main crustal forming processes for these ophiolites was \( \sim 120–125 \text{ Ma} \) (review in Hébert et al., 2012), contemporaneous with the Gangdese arc magmatism. In the classical model (scenario 2) these zircons are considered to represent the fore-arc of the Gangdese arc (e.g., Xia et al., 2003). This interpretation is however at odds with several geological observations: (1) many of these zircons are associated with deep sea shales, limestones, and radiolarians indicating that at least some of these rocks formed far from any continental source (e.g. Zibavev et al., 2003). (2) Different ophiolitic bodies are considered to represent fore-arc (e.g., Luobuza), intra-oceanic arc (e.g., Jinlu) and back-arc (e.g., Yungbu), constituting a complete supra-subduction zone system (Dubois-Cote et al., 2005; Hébert et al., 2012). If these rocks are indeed associated with the same subduction system as the Gangdese arc, that would indicate that there would have been a second volcanic chain in the fore-arc, separated by a “back arc” basin from the main continental arc. No currently active continental margin subduction zone system shows such a double volcanic chain; (3) finally, palaeomagnetic data suggest that these ophiolites were in an equatorial position in the mid Cretaceous (Abrajevitch et al., 2005), approximately at the same latitude as the KLA (Khan et al., 2009), but \( \sim 2500 \text{ km} \) south of the Gangdese arc that formed at \( \sim 25 \text{ N} \) (Dupont-Nivet et al., 2010; Tan et al., 2010; van Hinsbergen et al., 2012).

### 5.4.2. Detrital zircon record

The appearance of Eurasian derived detritus (e.g., detrital zircon grains) within the Tethyan Himalayan series is used to constrain the timing of the India–Eurasia collision. In this approach, the mineralogy of the sediments is used to constrain the source regions, and recent studies have focused on a pronounced shift in U–Pb age spectra of the detrital zircons at the Ypresian. Whereas pre-Ypresian samples show a rather simple age spectra dominated by a \( \sim 120 \text{ Ma} \) age peak, post-Ypresian rocks show an age spectra characterized by a peak at \( \sim 55–60 \text{ Ma} \) and a broad double peak at \( \sim 90–100 \text{ Ma} \) (Fig. 7). The source of these zircons grains is actively debated: (1) one model considers these zircon crystals to be derived from the intra-oceanic arc (e.g., Aitchison et al., 2011), in accordance with our proposed scenario 1 above; whereas, (2) others consider these zircons to be derived from Eurasia (e.g., Najman et al., 2010 the scenario 2; Wang et al., 2011). The previous studies are mainly based on apparent matches between age peaks observed in the Gangdese arc, the ophiolitic remnants, and in the detrital zircon age spectra. Because the dataset for the arc was very limited prior to our study, and that the extent and origin of the sedimentary units is still disputed (Aitchison et al., 2011; Clift et al., 2002; Henderson et al., 2011; Wu et al., 2010), we re-evaluate the two possible scenarios by comparing in Fig. 7, the detrital zircon record from pre- and post-Ypresian Tethyan sediments with the igneous zircon record of the Eurasian Margin, and our new data from the KLA and other arc fragments found within and south of the Tsango suture zone.

Probably the most obvious conclusion from Fig. 7 is that neither the data set from the Eurasian margin, nor the intra-oceanic arc system, fully matches the detrital zircon age record. The 55–60 Ma peak in the post-Ypresian samples is present in both the oceanic arc and continental margin dataset, thus inconclusive. A lack of zircons dates \( \sim 75 \text{ Ma} \) in the continental margin dataset could mirror the few zircon crystals of that age in the post-Ypresian samples. The peaks at \( \sim 90–100 \text{ Ma} \) in the detrital zircon are very well represented in the intra-oceanic dataset but not recorded in the continental margin.

The correlation of the detrital U–Pb zircon record to igneous activity recorded in the source region is complicated. For example, the preservation likelihood of less mature oceanic arc systems is low and it is possible that part of the source region of the zircon grains were subducted together with the leading edge of India during the final India–Eurasia collision (Boutelier and Chemenda, 2011; Boutelier et al., 2003; Burg, 2006). Additionally, it is in general difficult to compare igneous U–Pb zircon dates to detrital U–Pb zircon dates because it is unknown if age peaks are truly related to large volumes of temporally constrained igneous pulses. Finally, age peaks could be influenced by the analytical method used which rely on significantly different number of analyzes, possibly resulting in a strong bias in “age peaks” toward analyses done by LA-ICPMS. Altogether, this indicates that an argumentation based on the comparison of age peaks between the Eurasian margin, the arc, and the detrital zircon alone, does not allow deciphering the two main geodynamic scenarios.

### 5.4.3. Change from localized calc-alkaline magmatism to widespread alkaline magmatism

The continental margin calc-alkaline magmatism recorded in the Lhasa terrane, that formed the Gangdese batholith, as well as the Linzizong volcanics, ended at \( \sim 40 \text{ Ma} \) (e.g., Wen et al., 2008).
Subsequently, the magmatic activity switched to widespread alkali-dominated intra-plate magmatism distributed throughout the Tibetan plateau (Ding et al., 2007). This change in the location and chemical characteristics of magmatism at ~40 Ma is in the Tibetan Plateau is readily explained by the geodynamic scenario 1 above. In this model the calk-alkaline magmatism is related to oceanic subduction, whereas widespread alkali magmatism is postcollisional in accordance with the frequent interpretation of alkaline magmatism worldwide. However, in the case of tectonic scenario 2 above, the switch in magmatism at ~40 Ma can only be poorly explained. For example, an ad hoc slab break-off model, where the oceanic slab detached from the subducting Indian continental lithosphere right after the onset of collision, has been proposed to explain the presence of the ~50–40 Ma mantle-derived melts with subduction characteristics (e.g. Wen et al., 2008). Yet, numerical model results indicate that slab break-off is unlikely to occur simultaneously with collision (van Hunen and Allen, 2011) and the magmatic record where slab tears and break-offs currently occurs, e.g., in the Mediterranean region, is generally highly diverse and not “simple” calk-alkaline melts.

Based on these observations and the fact that both the Gangdese and Karakoram batholith record subduction related magmatic activity until ~40 Ma (Searle et al., 2010; Wen et al., 2008), we favor a geodynamic scenario involving two subduction zone systems (scenario 1; Fig. 6) as the one most compatible with our current understanding of the broader geology (e.g. Aitchison and Davis, 2004). We speculate that the KLA and the ophiolitic fragments of the Tsango were part of the same intraoceanic arc chain present in the Neo-Tethys south of the Eurasian margin. The multiple low velocity zones observed in the upper mantle and interpreted as slab remnants could reflect these two subduction systems (Haffkenscheid et al., 2006; Replumaz et al., 2010; Van der Voo et al., 1999). In this model the extension towards the west of what we call the Shyok Sea, between the two subduction systems, could be the Cretaceous–Miocene Pishin sedimentary belt (Khan et al., 2012) that have been observed behind the ophiolites in the western extension of the KLA (Kakar et al., 2012).

Our scenario implies that two suture zones may be present in the central and eastern Himalaya that have not been fully identified so far. This could stem from the fact that both suture zones are now represented as the apparent single Tsango suture, that often show duplications (e.g. Gansser, 1980), and are not yet resolved. Alternatively, an unidentified major suture zone exits north or south of the Tsango: either within the Lhasa Terrane, but there are no evidence for a major fault zone and/or tertiary oceanic sediments; or, included in the poorly understood Tibetan Himalaya sequence that is composed of highly variable sedimentary units of different origin (e.g. Burg and Chen, 1984).

6. Conclusions

We have presented results used to reconstruct the temporal and spatial isotopic record of the Kohistan–Ladakh Paleo-Island Arc granitoids, and showed that India collided first with the Arc at 50.2 ± 1.5 Ma and that the final India–Arc–Eurasia collision occurred at 40.4 ± 1.3 Ma in northwestern Himalaya. The geological record of the whole belt is in accordance with a scenario of a final India–Eurasia collision at ~40 Ma. Considering a post 40 Ma Indian plate velocity of 50 mm per annum (Molnar and Stock, 2009), the total length of the Indian lithosphere accommodated in the orogen since the final collision is ~2000 km. This estimate is ~1000–1500 km less than the convergence that would have had to be accommodated if collision occurred at 50 Ma (e.g., ~3000–3500 km). A total postcollision India–Eurasia convergence of ~2000 km resolves a long-lasting debate that pertains to the importance of different deformation processes that accommodated convergence (e.g., Molnar and Tapponnier, 1975; Roeden et al., 2008; Tapponnier et al., 1982). Estimates on the shortening recorded in the Himalaya and Eurasia (DeCelles et al., 2002; van Hinsbergen et al., 2011) are broadly consistent with the reduced amount of convergence due to our 40 Ma collision scenario.

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Appendix A. Supplementary materials

Supplementary data associated with this article can be found in the online version at http://dx.doi.org/10.1016/j.epsl.2013.01.023.

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