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Abstract: This geochemical, geochronological and structural study of intrusive rocks in the Sakya Dome of southern Tibet has identified two distinct suites of anatectic granites that carry contrasting implications for the tectonic evolution of the India-Asia collision zone. The northern margin of the dome core was intruded by anastomosing, equigranular two-mica garnet granites between 28.1 ± 0.4 Ma and 22.6 ± 0.4 Ma, coeval with top-to-the-south shear. Trace-element and isotopic (Sr-Nd) characteristics indicate an origin from partial melting of a biotite-bearing source in the Indian crust, under conditions of high fluid-phase activity. These granites thus provide evidence for the melt weakening required by some thermo-mechanical models that predict the southwards extrusion of a low-viscosity channel during the Oligocene. Evidence for subsequent shear-sense reversal may document initiation of this process. However, a younger suite of porphyritic two-mica granite plutons, emplaced between 14.5 ± 0.9 Ma and 8.81 ± 0.22 Ma, are derived from anatexis of muscovite-bearing metasediments of the High Himalayan Series under fluid-absent conditions. Ar-Ar cooling ages of 14.4 to 8.0 Ma from the Sakya dome postdate crystallisation of the Oligocene granite suite by ca. 10 Ma, but are coincident with mid-Miocene granite emplacement, suggesting uplift to depths of <10 km by the mid-Miocene. We propose that plate flexural response to Miocene slab steepening is a likely cause of dome uplift, and that this exhumation of mid-crustal rocks triggered decompression melting at 15-9 Ma and emplacement of discrete granite plutons into the upper crust under brittle conditions.

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THE CONTRIBUTION OF CRUSTAL ANATEXIS TO THE TECTONIC EVOLUTION OF INDIAN CRUST BENEATH SOUTHERN TIBET

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ABSTRACT

This geochemical, geochronological and structural study of intrusive rocks in the Sakya Dome of southern Tibet has identified two distinct suites of anatectic granites that carry contrasting implications for the tectonic evolution of the India-Asia collision zone. The northern margin of the dome core was intruded by anastomosing,

equigranular two-mica garnet granites between 28.1 ± 0.4 Ma and 22.6 ± 0.4 Ma, coeval with top-to-the-south shear. Trace-element and isotopic (Sr-Nd) characteristics indicate an origin from partial melting of a biotite-bearing source in the Indian crust, under conditions of high fluid-phase activity. These granites thus provide evidence for the melt weakening required by some thermo-mechanical models that predict the southwards extrusion of a low-viscosity channel during the Oligocene. Evidence for subsequent shear-sense reversal may document initiation of this process. However, a younger suite of porphyritic two-mica granite plutons, emplaced between 14.5 ± 0.9 Ma and 8.81 ± 0.22 Ma, are derived from anatexis of muscovite-bearing metasediments of the High Himalayan Series under fluid-absent conditions. Ar-Ar cooling ages of 14.4 to 8.0 Ma from the Sakya dome postdate crystallisation of the Oligocene granite suite by *ca.* 10 Ma, but are coincident with mid-Miocene granite emplacement, suggesting uplift to depths of <10 km by the mid-Miocene. We propose that plate flexural response to Miocene slab steepening is a likely cause of dome uplift, and that this exhumation of mid-crustal rocks triggered decompression melting at 15-9 Ma and emplacement of discrete granite plutons into the upper crust under brittle conditions.

246 words

Keywords: **Himalaya, gneiss domes, granite, isotope geochemistry, channel flow**

INTRODUCTION

The Himalayan orogen has resulted from the geologically recent collision of the continental lithosphere from two converging plates and provides a natural laboratory for studying the mechanisms responsible for orogenic evolution. The high-grade metamorphic rocks of the subducted Indian Continent, termed the High Himalayan Series, form the core of the orogen and their outcrop marks the prominent highlands of the Central Himalayan topographic front. Although critical to our understanding of the evolution of mountain belts, the mechanism by which these lithologies have been emplaced remains controversial (see Harrison, 2006 and Harris, 2007, and references therein). The classic study by Heim & Gansser (1939) of the metamorphic rocks of the High Himalaya describes them collectively as 'an enormous deep-rooted body of injected crystalline rocks, 10-20km thick', and implies that "sliding" (thrusting) took place at depth, not along a discrete plane but rather over a ductile zone. Competing models for the mechanisms that exhumed these rocks may be broadly classified under two main groupings, (i) coherent exhumation and (ii) exhumation by flow. The first group of orogenic models was inspired by Burchfiel et al. (1992) who recognised that the High Himalayan Series formed a wedge, tapering at depth to the north and bounded below by the Main Central thrust and above by the South Tibetan detachment. Applying the work of Dahlen (1990) on Critical Wedge theory to the Himalayan orogen, a series of authors including Harris & Massey (1994) and Avouac (2003), developed a model in which a wedge of mid-crustal rocks is exhumed as a coherent body. Recently, tectonic wedging where the crustal wedge is bounded up-dip

(to the south) by the merging of the Main Central thrust and the South Tibetan detachment has been proposed in several studies (Yin, 2006; Webb et al., 2007; Webb & Yin 2008). Metamorphic studies from the Nepal Himalaya have provided evidence that P-T gradients could be readily explained by successive underplating of an evolving wedge (Kohn, 2008).

The second group comprises thermomechanical extrusion models that envisage the ductile transport of a 'pipe/sheet to the surface' of low viscosity mid-crustal rocks (Burchfiel & Royden 1985, Nelson et al., 1996; Beaumont et al. 2001, 2004; Grujic et al., 2002; Jamieson et al., 2004) implying the decoupling of a ductile lower or mid crustal segment or channel from both the upper crust and the underlying mantle (Royden et al., 1997). An essential requirement of the ductile 'channel' flow models is a reduced bulk viscosity of rocks that define the channel, such as would be induced by the generation of a small (< 7%) melt fraction (Hollister and Crawford, 1986; Rosenberg and Handy, 2005), a process termed 'melt weakening' (Beaumont et al., 2001, Grujic et al., 2002). The Miocene leucogranites that intrude the High Himalaya are unequivocally sourced from the melting of pelitic lithologies from within the metamorphic pile (Harris & Massey, 1994) and so provide a potential mechanism for effecting viscosity reduction. Seismic and magnetotelluric surveys of southern Tibet and the north Himalaya have identified anomalous geophysical characteristics — 'bright' low-velocity reflectors — that can be interpreted as a zone of present-day partial melting at depths of 15-20 km beneath a north-south rift zone (Makovsky et al., 1999; Nelson et al., 1996; Unsworth et al., 2005) and 20-40 km beneath the NW Himalaya (Caldwell et al., 2009, *in press*). However, these bright reflectors have also been interpreted as the presence of an aqueous fluid phase (Makovsky & Klemperer,

1999; Li et al. 2003), and there remain uncertainties in the geophysical evidence for mid-crustal melting on a regional scale (Harrison, 2006). Therefore, geological evidence is now required to establish the timing, distribution and mechanisms of crustal melting to assess the possible role of melt weakening within the crust during continental collision. Models that seek to link extrusion of the mid-crust with crustal melting make specific and testable predictions about the timing of crustal melting. Hence this contribution investigates the geochemistry, chronology and structural setting of anatectic granites from southern Tibet with a view to assessing the viability of melt weakening in relation to extrusion models for the Himalayan orogen.

Evidence for Tertiary crustal melting is found in two tectonic environments (Debon et al. 1986). Firstly, peraluminous leucogranites intruding and derived from the High Himalayan Series are exposed across the High Himalaya, and generally range in age from 23 Ma to 17 Ma (Harrison et al., 1998) with mid-Miocene ages (as recent as 12 Ma) restricted to the eastern Himalaya (Daniel et al., 2003; Edwards and Harrison, 1997; Schärer et al., 1986). Secondly, north of the High Himalaya a belt of leucogranites intrude High Himalayan Series rocks exposed in the core of the North Himalayan antiform, an east-west trending structure located *ca.* 50 kilometres south of the Indus-Yarlung-Tsangpo suture zone (Hauck et al., 1988; Hodges 2000; Watts et al., 2005). Published ages for these North Himalayan granites range in age from 28 Ma to 9 Ma (Aoya et al., 2005; Burg et al., 1984; Chen et al., 1990; Lee et al., 2000, 2004, 2006; Quigley et al., 2006, 2008; Schärer et al. 1986; Zhang et al., 2004) with some evidence for migmatite formation as early as 35 Ma (Lee and Whitehouse, 2007). Whereas the High Himalayan granite suite is related to fluid-absent melting of the High Himalayan Series source region during exhumation (Harris et al., 2004), the

North Himalayan granites span a much wider age range, encompassing higher solidus temperatures and deeper source regions (Zhang et al., 2004), potentially providing evidence for the melt weakening required by the channel-flow model when return (i.e. southward) flow was initiated during the Oligocene (Jamieson et al., 2004). Thus, the key to understanding the relationship between crustal anatexis and the regional tectonic evolution of the Tibet-Himalayan orogen lies in unravelling the chronology and melt mechanisms of granite formed within the gneiss domes of the North Himalayan antiform. This study primarily comprises a detailed geochemical and geochronological analysis of eight granitic bodies intruding the Sakya¹ gneiss dome, building on previous work in the region (Schärer et al., 1986; Lee et al., 2004; Zhang et al., 2004; Watts et al., 2005; King et al., 2007). We present major and trace-element abundances, Sr-Nd isotope data and U-Pb accessory phase ages for the intrusives, together with Ar-Ar cooling histories. The structural context is based on the detailed mapping of Lee et al. (2004), and we include some additional structural observations from areas not visited by these authors, notably from critical localities at the northern and south-western dome margins. We identify two distinct periods of anatexis that can be resolved in terms of their chronology, mechanisms of anatexis and structural settings. These anatectic periods document the changing thermal and tectonic conditions that prevailed within the deep crust of the northern Indian plate from the

¹ The Sakya dome is also in part referred to as the Mabja dome by Lee et al., (2004). However, in this study we prefer the term Sakya Dome, after the major town in the area, in reference to the whole outcrop depicted in Figure 2, and thus avoiding confusion with the Mabja (also spelled Maja) granite (Schärer et al. 1986).

Oligocene to the late Miocene and so constrain the potential role of crustal anatexis in facilitating any possible southwards crustal flow during the post-collisional convergence of the Indian and Asian plates.

REGIONAL SETTING OF THE GRANITES

The east-west trending North Himalayan antiformal structure crops out within the Tethyan Himalaya and exposes a discontinuous alignment of granite gneiss domes (Fig. 1) that can be traced along strike for over 500 km (Le Fort et al., 1987; Burchfiel et al., 1992; Hodges 2000; Watts et al., 2005). The cores of the domes comprise variable proportions of Paleozoic granitoid gneisses, and Neoproterozoic to early Paleozoic high-grade metasediments including migmatites (Chen et al., 1990 (Kangmar), Lee et al., 2000, 2004 (Kangmar and 'Mabja' Domes, respectively), Zhang et al., 2004 (Sakya Dome) and Quigley et al., 2006, 2008 (Kampa Dome)) with petrological and isotopic characteristics that indicate they represent tectonic windows of High Himalayan Series crust (Zhang et al., 2004). To a minor extent, components of Lesser Himalayan Series crust have also been identified in core gneisses of the Gurla Mandhata dome (Murphy, 2007) at the western extremity of the North Himalayan antiform. Tertiary peraluminous granites are also exposed along the antiformal axis. In contrast to the High Himalayan Granites, these Tertiary granites, termed the North Himalayan granites, intrude both the gneiss dome core lithologies and the metasediments belonging to the Tethyan Sedimentary Series of largely Triassic age that mantles the domes. The upper bounding contact between the dome core gneisses and the mantling Tethyan metasediments is a kinematically complex

high-strain zone that has been correlated with the Miocene South Tibetan detachment, or, according to some studies, its northern continuance, known as the Great Counter thrust (Heim and Gansser 1939; Yin et al., 1994; Yin 2006, Webb et al., 2007). From east to west the domes and their associated granites have been termed the Kangmar, Kampa (intruded by the Kampa granite), Sakya (intruded by the Kuday, Wing, Kua, Lijun, Donggong, Gomdre, Kouwu and Mabja granites), Lhagoi Kangri, Malashan (intruded by the Cuobu, Malashan and Paiku granites), Gurla Mandhata and Leo Pargil (NW Himalaya) gneiss domes (Aoya et al., 2005; Burg et al., 1984; Chen et al., 1990; King et al., 2007; Lee et al., 2000, 2004, 2006; Murphy et al., 2002; Murphy, 2007; Quigley et al., 2006, 2008; Schärer et al. 1986; Thiede et al., 2006; Zhang et al., 2004).

Key to unravelling the regional significance of the Tertiary melts is the timing of intrusion relative to dome structural development and uplift. The formation of the antiform itself remains enigmatic and has been variously explained as (i) a consequence of imbrication associated with wedge tectonics (Burg et al., 1984; Ratschbacher et al., 1994), (ii) duplex development above a thrust ramp (Burg et al., 1984; Lee et al. 2000, 2006 and Yin 2006), (iii) underthrusting by or ramping within a low-viscosity crustal channel (Beaumont et al., 2001), (iv) extensional tectonics associated with the South Tibetan detachment system (Chen et al., 1990), (v) passive collapse associated with diapirism (Aoya et al., 2005; Lee et al., 2004; Kawakami et al., 2007; Teyssier and Whitney, 2002) or gravitational instability (Le Fort et al., 1987), and (vi) upwelling of ductile lower crust under horizontal compression (Yin et al., 1999). Cooling histories for the Kangmar and Mabja¹ domes are broadly consistent with mid-Miocene dome formation (Maluski et al., 1988; Lee et al., 2000)

related to motion on the South Tibetan detachment or the associated thrust ramp systems.

Competing hypotheses for the origin of the Tertiary melts include shear heating along the flat section of a continuously active décollement (Harrison et al., 1998) and decompression melting of the mid-crust during exhumation and doming (Lee et al., 2004; Zhang et al., 2004).

This present contribution is a study of crustal anatexis within the Sakya dome which, covering an area approximately 50km x 50km (Watts et al., 2005), arguably represents the largest and most complex of the central gneiss domes.

Our field studies have determined that the Paleozoic granitoid gneisses forming the core of the Sakya dome, together with their Tethyan Sedimentary Series cover, are intruded by Tertiary granite melts that define two distinct groups: (i) anastomosing, equigranular, leucogranites (termed AEG) and (ii) discrete medium to coarse grained porphyritic leucogranite plutons (termed DPP). These two groups are distinguishable by their petrology, geochemistry and their emplacement geometries.

PETROLOGY AND STRUCTURAL RELATIONSHIPS OF THE GRANITES

Anastomosing, equigranular granites (AEG).

Granites of this type include the intrusives at Kunday, Wing, Kua and Lijun (Fig. 2), which form sheets, lenses and dikes with irregular contacts that are generally parallel to the schistosity of the country rock. They are characterised by a medium-grained, locally foliated texture and comprise quartz (~35%), oligoclase (~37%) with significant myrmekitic intergrowth and perthitic microcline (~25%); minor phases

include muscovite, biotite, garnet and tourmaline. Accessory phases include zircon (found in all bodies), xenotime (Kuday, Kua), monazite (Kua) and allanite (Lijun).

The field relations of this group of granites are typified by good exposures along the northern contact of the Sakya gneiss dome at Kuday (Figs. 2 and 3) which is characterised by a sharp sheared boundary between foliated granitic gneisses of the dome core (referred to here as the Kuday gneiss complex) and staurolite-grade garnetiferous schists of the Tethyan sedimentary cover. Within the granitic gneiss core, multiple anastomosing generations of AEG melt emplacement are evident, from undeformed AEG bodies (0.5 m dikes to >10 m sheets) to earlier foliated generations (Fig. 4a, b). The Kuday gneiss complex comprises predominantly biotite augengneisses containing alkali feldspar augen (≤ 3 cm) + quartz + biotite + plagioclase + muscovite + garnet $\pm$ sillimanite, and may be correlated with the Cambrian gneisses of the Kangmar dome (Lee et al., 2001, 2004; Zhang et al., 2004; unit OG of Lee et al., 2004). Down section (south from Kuday, toward the centre of the dome) there is a marked increase in biotite content within the foliated AEGs. This trend is accompanied by increasing abundance of xenolithic bodies of schistose and mylonitic material (Fig. 4a and b). Large, dark pods on the order of 2-5 m across appear to be biotite-rich restite pods as opposed to a more mafic melt generation (Fig. 4b). These observations are similar to those made by Lee et al. (2004), in the western and southern parts of the dome where the Kua and Donggong granites intrude the unit POP, identified by these authors as Paleozoic orthogneisses and metasedimentary rocks. Diffuse mutual margins between foliated and unfoliated AEG bodies have been observed in both the Kua and the Kuday area. All internal dome core structures are ubiquitously cut by AEG pegmatites. In the foliated facies, the main foliation (S_2),

defined by planar mica alignment, largely defines a domal geometry, dipping west, north and east radially away from sample point JK3/05, ~4 km south of Kuday village (Fig. 2). Foliation S_2 defining the overall domal geometry is taken up by early AEG bodies but not by later generations. This tectonic fabric is a ductile, pervasive foliation, revealed as a crenulation of an earlier fabric (S_1) in lower-strain Tethyan metasediments at some distance from the gneiss contact. This relationship between foliations in the schists agrees with the observations of Lee et al. (2004) elsewhere in the Sakya dome.

Large (0.5 to 2 m thick) AEG dikes are also observed to intrude the Tethyan Sedimentary Series at distances up to 200 m above the contact (Fig. 4 c and d). Although we were unable to trace major individual units at accessible exposures across the contact into the metasediments of the Tethyan Sedimentary Series, abundant small (~50 mm) late-stage pegmatitic AEG dikes traverse the contact into the metasediments (Fig. 5 a and b). It is important to note that the AEG dikes are distinct in orientation and composition from the north-trending dacite dikes that intrude the metasediments of the Tethyan Sedimentary Series farther up section (Fig. 4 c and d), cross-cutting all lithologies (King et al., 2007).

The AEG dikes above the contact truncate the schistose foliation (S_2 dipping 20-30° North) at a low angle and appear to have been strongly rotated in a dextral sense within the metasediments (Fig. 5c), assuming vertical emplacement, equating to top-to-south shear. Up-section, the dikes become pinched, boudinaged and progressively more foliated before becoming largely absent ~200 m from the contact. Major-element analysis of a foliated AEG dike (JK3/13b, Table 1), recovered some 50 m above the contact, indicates that it is of similar composition to the undeformed granite

sampled from the gneiss-dome core, dated at 27.5 ± 0.5 Ma (T100, Zhang et al. 2004). Rotation and intense shearing of these dikes during syn-deformational emplacement makes it impossible to trace any single dike directly back to the more massive facies leucogranite bodies intruding the dome core lithologies. The Cambro-Ordovician granitic gneiss facies, characteristic of the cores of the North Himalayan domes, was not observed anywhere above the AEG/ Tethyan Sedimentary Series contact.

Along the granite-metasediment contact, schists of the Tethyan Sedimentary Series are strongly deformed; garnets, up to 30 mm across, gradually decrease in size and abundance up section as strain increases. Approximately 200 m above the contact there is a garnet-free mylonitic zone of calc-silicate, amphibolite and marble (Fig. 5d). This mylonitic zone also marks the upper structural limit of AEG dike outcrop.

Although most shear sense indicators post-date the main ductile foliation, making many early potential kinematic indicators ambiguous, garnet pressure shadows and sigmoidal quartz augen both appear indicative of top-to-south shear within 10 m of the contact. Moreover, many small (<200 mm thick), AEG pegmatitic dikes that crosscut the main ductile foliation and AEG dike/Tethyan Sedimentary Series zone are weakly sheared top-to-south within the schistose lithologies (Fig. 5 a & b). We therefore interpret the field observations at the contact between the granitic gneiss dome cores and their TSS cover as indicative of top-to-south ductile shear (D_2), accompanied by the progressive emplacement of AEG melts.

The top-to-south shear is associated with formation of a fully ductile, pervasive foliation (S_2). However, from about 5 m above the granite/metasediment contact, this fabric is overprinted by widespread shear bands (S_3) spaced a few centimetres apart, with C' shear surfaces that deflect the S_2 foliation to indicate a top-to-north shear

sense (Fig. 6a). These S_3 shear bands are locally associated with shear fractures that displace quartz veins and the Miocene dacite dykes in the same sense as the ductile shear bands, hence exhibiting behaviour intermediate between ductile and brittle that probably occurred at lower grades/higher crustal levels than formation of the main S_2 foliation. The association of the S_3 shear bands with chlorite-bearing pressure shadows adjacent to garnet is also indicative of lower grades for the top-to-north deformation (Fig 5 e and f; 6d). This top-to-north overprint increases in intensity up section to become the dominant shear indicator in the mylonitic zone (foliation 26/352 and lineation 20/000) (Fig. 5d). When considered with the increase in strain of the AEG dikes, we conclude that partitioning of strain at lower grade into the schists above the contact was coincident with a relative reversal in shear sense (D_3).

Discrete porphyritic plutons (DPP).

Granites from this group include intrusions from Donggong, Gomdre, Kouwu and Mabja (Fig. 2) that form discrete plutons with steep contacts cross-cutting country-rock fabrics and layering; their elliptical outcrop pattern ranges in size from 3 to 10 km across. The plutons comprise quartz (~36%), oligoclase (~37%), and perthitic microcline (~24%) forming phenocrysts up to 100 mm across; additional phases include muscovite, biotite and tourmaline. Accessory phases include zircon and monazite (in all bodies), and xenotime (Gomdre, Mabja). Myrmekitic intergrowth is notably absent from the DDP granites.

At Gomdre La (28°43.15N, 88°01.65E; near sample site JK4/13g, Fig. 2), Tethyan sediments with top-to-south shear bands overlie sheared Gomdre granite on a brittle detachment. Large (~100 mm) feldspar augen wrapped by the ductile mica fabric in

the granite also indicate top-to-south shear, and this shear fabric dies out within 10 m of the detachment. We interpret these localized fabrics to reflect strain during buoyancy-driven granite emplacement (D_4).

These D_4 structures are transected by major brittle north-south striking normal faults (D_5) with DPP granites in their footwalls at Gomdre and Mabja. Major north-south striking normal faults also place DPP granites in their footwalls at Kouwu and Donggong. Numerous pseudotachylite veins cutting the footwall granites are spatially associated with these normal faults.

Metamorphic History

Thin sections from the Tethyan metapelitic samples taken from Kunday, Lijun and Mabja complement sampling by Lee et al., (2004), for the Kouwu/Donggong area of the Sakya¹ dome. Petrographic examination reveals a series of isograds defined by prograde mineral assemblages, with metamorphic grade increasing toward the axis of the antiform. These isograds are roughly concentric to the domal structures defined by the warped stratigraphy, and parallel the lithologic contacts and the S_2 foliation (Fig. 2).

The lowest grade schists are characterized by abundant muscovite and chlorite, and at somewhat deeper structural levels, prismatic chloritoid is associated with muscovite and chlorite, defining the chloritoid zone. The relationship between garnet and chloritoid isograds is not clear-cut since rocks containing garnet $\pm$ chloritoid $\pm$ chlorite + muscovite + plagioclase + quartz assemblages are commonly intercalated with garnet-absent, chloritoid-bearing assemblages, depending on the alumina content

of the protolith. Garnet porphyroblasts are generally sub- to euhedral 1-20 mm size grains.

Down section, the breakdown of chloritoid in the presence of biotite yields garnet + staurolite + chlorite $\pm$ biotite, defining the staurolite zone; chloritoid and biotite commonly form inclusions in garnet. At deeper structural levels, garnet + chlorite assemblages reacted to form staurolite + biotite $\pm$ kyanite (as rare inclusions in garnet) in some areas. In this zone, euhedral to subhedral staurolite porphyroblasts of 0.5-1mm size are closely associated with pre-deformational garnet that is breaking down to biotite. No higher grades are evident in the metapelites of the Tethyan series rocks at Kuday since the series is truncated at lower structural levels by the sheared contact with the Kuday granitic gneisses and AEG intrusives. However, kyanite has been reported previously in the Kuday granite below the contact, by Zhang et al., (2004). Microstructures reveal the relative age relations between prograde metamorphic porphyroblast growth, which we refer to as M_1 consistent with Lee et al., 2004, and the development of the S_2 foliation.

(1) Chloritoid porphyroblasts contain straight inclusion trails defined by fine-grained quartz + white mica $\pm$ biotite $\pm$ chlorite that are oblique to the external foliation, S_2 . The S_2 foliation also wraps weakly around the chloritoid porphyroblasts leading to poorly-developed strain shadows (Fig. 7b). These microstructural relations indicate typical pre- to syn- S_2 foliation porphyroblast growth.

(2) Garnet porphyroblasts exhibit a range of microstructures. Porphyroblasts within the garnet zone contain straight to gently warping microfolded inclusion trails defined by quartz, micas, and Fe-oxides that are often discontinuous with the external foliation (S_2), implying garnet growth before S_2 development. However, some grains

appear syn-kinematic, with lightly developed snowball inclusion trains that are continuous with the external S_2 fabric. The S_2 foliation also wraps around garnet porphyroblasts leading to the development of quartz-filled pressure shadows (Fig. 7a).

(3) Garnet porphyroblasts within the staurolite zone lack well-defined inclusion trails and most appear partially altered to biotite. S_2 foliation wraps around these porphyroblasts, producing strain shadows (Fig. 7c). These textures suggest that garnet growth within the garnet zone was both pre- and syn-tectonic with D_2 deformation, whereas garnet growth predated D_2 deformation in the staurolite zone.

(4) Staurolite porphyroblast range in habit from small, anhedral grains to large poikiloblasts. Medium-sized staurolite porphyroblasts locally contain inclusion trails defined by quartz and Fe-oxides that are continuous with the external S_2 foliation, whereas the S_2 foliation wraps moderately around large staurolite porphyroblasts, creating strain shadows (Fig. 7c). These fabrics suggest that staurolite growth was syntectonic with D_2 deformation.

(5) Chlorite aggregates form in the pressure shadows of S_2 garnet porphyroblasts and exhibit both continuous and discontinuous growth habit relative to the S_2 fabric (Fig. 7d), suggesting that this mineral grew late syn- to post- D_2 deformation.

In summary, moderate to high-grade metamorphic rocks observed in the Kудay, Lijun and Mabja Tethyan metapelitic series define a Barrovian metamorphic facies series, and exhibit peak metamorphic conditions syn-kinematic with high-strain D_2 deformation, with limited retrogression, syn- to post-high strain D_2 deformation.

ANALYTICAL RESULTS

Additional detailed data tables are contained in the related Data Repository files[†].

Elemental geochemistry

The major-element abundances of the granites intruding the Sakya dome (Table 1) indicate peraluminous compositions that are broadly similar to the leucogranites of the High Himalaya (see Ayres and Harris, 1997 and references therein). However, the two distinct granite groups identified above from textural and field relations may also be distinguished by differing calcium abundances: for the AEGs, higher CaO (average 1.84 %, ranging from 0.77 – 2.05 %) are indicated by the Kuday, Kua, Wing and Lijun bodies, whereas the DPPs (Mabja, Donggong, Kouwu and Gomdre) show lower CaO (average 1.02 %, ranging from 0.73 – 1.55 %).

In general, trace-element variations of the North Himalayan granites from the Sakya Dome are large, overlapping with compositions typical of High Himalayan granites. The variable abundances of most trace elements in the North Himalayan granites obscure any systematic distinctions between the constituent groups. The notable exceptions are the HREE and Y which are distinctly higher in the AEG group. This distinction is exemplified by the chondrite-normalised REE plot (Fig. 8); whereas profiles from both granite groups show negative Eu anomalies, the AEG show lower La/Yb ratios through enrichment of REEs heavier than Dy. The trace-element distinction between the two-groups is best illustrated by the plot of Y against La/Yb (Fig. 9).

Those granites with highest Y and HREE (Kuday, Kua) contain the accessory phase xenotime, in addition to garnet. However, it is important to note that abundant xenotime is also present in the Gomdre pluton (from the DPP group) yet its compositions show no significant Y enrichment. This suggests that the presence of garnet, rather than xenotime, is the controlling factor for Y enrichment of the AEG.

Microprobe analysis of garnet separates from AEG assemblages (Table DR3) are spessartine-rich (almandine = 35-68%, pyrope <2%, spessartine = 19-36%, grossular = 11-41%). By contrast, garnets from the metapelitic assemblages within the Tethyan metasediments are depleted in both spessartine and grossular (almandine = 61-70%, pyrope = 7-11%, spessartine = 4-14%, grossular = 13-16%),

Zr and LREE abundances in granite melts each provide an independent constraint on solidus temperatures in the melt, based on zircon and monazite dissolution respectively (Watson and Harrison, 1983; Montel, 1993), the underlying assumptions being; (i) there is sufficient accessory phase to saturate the melt during its formation and (ii) the component mass of inherited zircon, or monazite, is insignificant. The most probable cause of error in the thermometry lies in the inheritance of zircon in the melt which can be tested by rejecting temperatures derived from zircon thermometry that are systematically higher than those derived from monazite thermometry (Ayres and Harris, 1997). Accessory mineral thermometry for the Sakya granites places approximate melt temperatures between 700 and 770 °C (Table 1). On average the AEGs formed at lower temperatures with a range of 700 to 730 °C, the exception being the Wing granite with somewhat higher temperatures of 745 °C (zircon) and 775 °C (monazite). The DPPs, in contrast, almost all yield slightly higher melt temperatures in the range 760 to 780 °C; the Mabja granite is an exception with a lower estimated melt temperature of 700 °C.

For the trace elements Rb, Sr, Ba that are predominantly sited in major phases in granitic rocks, Zhang et al. (2004) noted that the array defined by North Himalayan granite compositions could be compared with that of the High Himalayan granites in defining increasing Rb/Sr at decreasing Ba, indicative of vapour-absent incongruent

melting of mica from the source due to the large component of residual plagioclase and the formation of peritectic alkali feldspar (Harris and Inger, 1992). For the DPPs, as exemplified by the Mabja dataset, the trend of increasing Rb/Sr with decreasing Ba is confirmed, reaching Rb/Sr ratios as high as 20 at low Ba (12 ppm). However, for the AEGs, Rb/Sr ratios are found not to increase above values of *ca.* 5, even at lower Ba values (Table 1). This contrast suggests that whereas the evolution of the DPPs was controlled by low melt fractions resulting from fluid-absent melting conditions, the AEG bodies resulted from somewhat higher melt fractions. Given that the mineral thermometry indicates lower temperatures for the AEGs, we conclude that their formation was characterised by higher fluid-phase activities during melting.

Sr-Nd isotopes

Sr-Nd isotopic ratio analyses in bulk rock samples were undertaken to identify possible source regions for the melts. Potential sources include the High Himalayan Series and Lesser Himalayan Series of the Indian plate, and the Lhasa terrane of the Asian plate (Fig. 10a). All ratios are corrected to 10 Ma, the most recent time at which melt generation occurred. Data for the Asian plate derives from samples from the southern Lhasa terrane ranging from Gangdese intrusives and volcanics (high $\epsilon\text{Nd}(10)$ samples) to schists and gneisses of the basement (low $\epsilon\text{Nd}(10)$ samples).

The empirical fields, based on multiple sources cited in the caption to Fig. 10, demonstrate that whereas samples from the two Himalayan series of the Indian plate are distinct from those from the Lhasa terrane, the Tethyan Sedimentary Series field shows considerable overlap with both. It has long been recognised that the Indian

plate and the Lhasa terrane of the Asian plate were adjacent portions of Gondwana prior to the Jurassic (Dewey et al., 1988), and as reported recently by Dai et al., 2008, Mesozoic Tethyan Sedimentary Series rocks have transitional ϵ_{Nd} signatures which indicate source protoliths of both Indian and Asian origin, suggesting that the Lhasa terrane and Indian craton remained close enough during the Late Triassic to permit deposition of Tethyan sediments of Lhasa terrane derivation onto the northern margin of the Indian continent. However since the pre-Cenozoic (pre-collisional) isotopes, as well as Sm/Nd model ages (Dai et al., 2008), identify contrasting isotopic signatures for Indian and Asian crustal rocks, the available data suggest that end-Triassic rifting between them may have reactivated an ancient Gondwanan tectonic boundary that separated contrasting lithospheric domains or exotic terranes.

Analyses from the North Himalayan granites yield a range of $^{87}\text{Sr}/^{86}\text{Sr}$ from 0.7396 to 0.8757, and $\epsilon_{\text{Nd}}(10)$ from -7 to -20, suggesting a source within the High Himalayan lithologies of the Indian plate for all of these melts (Fig. 10a). The AEG bodies define a loose cluster indicating a fairly homogeneous source region. The DPPs cover a wider range, generally at lower ϵ_{Nd} values and extending (in the Mabja granite) to higher $^{87}\text{Sr}/^{86}\text{Sr}$ than the AEG cluster (Fig. 10b).

[†]GSA Data Repository item 2009##, [U-Pb accessory mineral isotopic data in Tables DR1 and DR2, and major-element garnet composition data in Table DR-3] is available online at <http://www.geosociety.org/pubs/drprint.htm>, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

Accessory phase dating

U-Th-Pb dating of zircon, monazite and/or xenotime was undertaken on 7 granitic samples from the Sakya dome. Due to the abundance of inherited zircon with Paleozoic cores in leucogranites of the Himalayan orogen, an adapted *in situ* U-Th-Pb dating methodology was essential for this study, details of which are given in the Appendix. Isotopic ratios are detailed in Data Repository Table DR1 (NuPlasma) & DR2 (Axiom data). The results (Figs. 11 and 12) are illustrated in either Concordia diagrams (incorporating a correction for common Pb where significant) or Tera-Wasserburg diagrams (where common Pb is not corrected but where the igneous age is determined by intercept with the Tera-Wasserburg Concordia curve). For the most part, data form concordant clusters, representing a single age component, or arrays of data that define a chord between two age components (where the laser analysis has sampled a mixture of zones with different ages). In some cases additional discordant analyses fail to fit a simple array and probably represent heterogeneous core age components within any given sample. Crystallisation ages are determined by either intercept method (in Tera-Wasserburg diagrams); intercept method of arrays on a Concordia diagram, or by calculation of Concordia ages, all using Isoplot3a software of Ludwig (2003). A summary of the U-Pb age data is presented in Table 2 (all errors are quoted at 2σ).

Kuday granite (AEG)

The complete Kuday dataset of sample JK3/05 comprises 65 conventional grain mount zircon analyses (Fig. 11a). These data define an eight-point concordant cluster with an age of 28.14 ± 0.44 Ma for the young component and a nine-point concordant

cluster with an age of 483.7 ± 6.8 Ma for the older component. These analyses were selected to represent a cluster of just one age component. Analyses that fall between these ages are consistent with differential mixtures of the two end-member components. The CL imaging of complex zoning and different age components within many of these grains (Fig. 11a) clearly supports this interpretation. The igneous external crystal shapes and/or igneous zoning in the outer rims of most of these crystals implies that the younger component was igneous, and we therefore interpret the ~ 28 Ma age as the crystallisation age of the Kuday granite. The crystallisation age of 28.14 ± 0.44 Ma is consistent with previously published crystallisation ages for the Kuday granite of 27.5 ± 0.5 Ma for zircon and 25.7 ± 0.7 Ma for xenotime, while the upper-intercept ages are consistent with Cambrian inheritance ages determined from this same sample (T100, Zhang et al. 2004) that was collected some 4 km north of the JK3/05 locality. Two individual grains indicate a second, much older, inheritance population with ages of ca. 900 Ma (Fig. 11a) probably reflecting mixed provenance within the source rock.

Wing granite (AEG)

The dataset for sample JK4/07a consists of 29 conventional grain mount zircon analyses and 31 flat mounted zircon analyses and the dataset is dominated by analyses falling between two end member components (Fig. 11b). A weighted mean of 7 concordant analyses at the upper intercept yields an age of 492.2 ± 4.6 Ma. Data from a selection of flat-mounted zircon grains indicates a discrete three-point concordant age of 27.46 ± 0.96 Ma, interpreted as the granite's crystallisation age.

Kua granite (AEG)

The Kua granite contains zircon, monazite and xenotime, all of which were conventionally mounted with analyses done on internal regions (Fig. 11c). Unfortunately, it was not possible to analyse very thin zircon rims for this sample. Five zircon analyses from the Kua granite cluster at 573.0 ± 24.0 Ma, the Concordia age of the three most precise concordant analyses. The two additional analyses with larger errors are shifted to the right of Concordia (i.e. normally discordant) with discordance and uncertainty being correlated, a situation indicative of common Pb for which an accurate correction has not or cannot be made. The intercept of all 6 points, when plotted on a Tera-Wasserburg diagram yield an age and uncertainty consistent with the above. The ~580 Ma zircon age is from inherited core regions of crystals with evidence of thin igneous rims. Analyses of conventionally mounted monazite (6 analyses) and xenotime (16 analyses) define age clusters at 23.7 ± 0.4 Ma (2 pt Concordia age) and 22.63 ± 0.37 Ma (Tera-Wasserburg intercept age), respectively. There is a clear difference in apparent age of monazite and xenotime, best explained by the presence of excess ^{206}Pb derived from initially incorporated excess ^{230}Th (Schärer 1984; Parrish, 1990) in Th-rich monazite. We conclude that the best estimate of crystallisation age is given by the 22.6 ± 0.4 Ma age of xenotime, within error of the published crystallisation age 23.1 ± 0.8 Ma for a pegmatite dike suite from the same area (Lee et al., 2006)

Lijun granite (AEG)

The zircon dataset comprises 22 conventional and 2 flat-mounted analyses of sample T03/33x (Fig. 11d). It shows a somewhat scattered array of data between

approximately 550 Ma and 25 Ma, with a single concordant analysis obtained from a flat-mounted zircon grain at 25.4 ± 6.0 Ma. Due to the thin igneous rims and the limited number of flat-mounted analyses, only one mid-Tertiary age was obtained, but this analysis is regarded as the best current estimate of crystallisation age. The scattered nature of the array indicates an inherited population of 500-550 Ma, but the sample also includes some grains that are even older, falling to the right of the main array. The ~25 Ma age is tentatively interpreted to be the best available crystallisation age of the granite.

T03/33x also contains accessory allanite. Eight analyses were normalised to Manangotry monazite, in order to attempt to manage probable excess Pb from ^{232}Th decay and/or high common Pb substitution for Ca in the allanite lattice. Common Pb corrected data plot on concordia with a lower intercept at 35.5 ± 8.8 Ma. However, as allanite is an extremely Th-rich mineral there is a high probability of excess Pb. Hence 35.5 Ma is a maximum age constraint.

Donggong granite (DPP)

The complete dataset, comprising 27 zircon analyses for sample JK4/12a (Fig. 12a) displays mixing of core and young rims, with a number of analyses clustering near the lower intercept of ~14 Ma. Four concordant analyses define an age of the inherited component of 473.3 ± 17.0 Ma. Many analyses define an array of mixed ages. Young zircon and monazite analyses all have common Pb but imprecise common Pb corrections, and therefore a Tera-Wasserburg plot of their data is used to define the age of the younger components. Ten zircons have both $^{206}\text{Pb}/^{238}\text{U}$ weighted mean and Tera-Wasserburg intercept ages of 14.4 ± 0.72 Ma and 14.5 ± 0.94 , respectively, and

the monazite analyses have a Tera-Wasserburg intercept of 15.0 ± 0.9 Ma. The likely presence of excess ^{206}Pb in monazite favours the zircon-derived age of 14.5 ± 0.9 Ma as the most reliable age of crystallisation. This is in broad agreement with a previous study that reported a zircon U/Pb ion-probe age for the Donggong granite of 16.2 ± 0.4 Ma (Lee and Whitehouse, 2007).

Gomdre granite (DPP)

Twelve analyses of conventionally mounted xenotime from sample JK4/13g measured on the Axiom MC-ICP-MS gave an age of 14.41 ± 0.62 Ma (12 point) (Fig. 12b) using a Tera-Wasserburg lower intercept calculated age. Analyses of flat mounted monazite (6 analyses) produced a very scattered cluster at 16.4 ± 2.5 Ma (Tera-Wasserburg intercept age), probably due to low signal voltages. Signal voltages for these flat mounted grains are, in general, one fifth of those from conventionally mounted grains, leading to the possibility of imprecise common-Pb corrections and poorly estimated errors. Again the difference in apparent age of monazite and xenotime is best explained by excess ^{206}Pb in monazite, implying that the 16.4 Ma estimate is a maximum age constraint. The best estimate for age of crystallisation is the Axiom xenotime analyses age of 14.41 ± 0.62 Ma. Thus crystallisation ages indicate that the Gomdre and Donggong granites are coeval, and within error of the age of the nearby Kouwu granite (14.4 ± 0.1 Ma, Zhang et al., 2004) determined from zircon and monazite dating.

Mabja granite (DPP)

Combined data from conventionally mounted monazite and xenotime grains from sample T03/25i yield two clusters of concordant data (Fig. 12c) with a four-point (3

xenotime plus 1 monazite) concordia age of 8.81 ± 0.22 Ma and a two-point xenotime concordia age of 11.20 ± 0.46 Ma taken from a 'core' sector of a single xenotime. The interpretation of the latter age is uncertain but could reflect either a xenocrystic ~11 Ma xenotime component or a mixture of ~9 Ma and older xenotime components in zoned crystals, representing a period of prolonged melt generation or emplacement. The consistency of the ~9 Ma group of analyses of both xenotime and monazite suggests strongly that it represents the best estimate for crystallisation of the Mabja granite. This age compares well with the published crystallisation age of 9.8 ± 0.7 Ma for the Mabja granite (Schärer et al., 1986).

Ar-Ar cooling ages

A north-south transect of 8 samples from the northern contact of the Kuday gneiss complex with the TSS towards the interior of the dome (Fig. 2) shows a largely flat profile of single-grain, spot ages on biotite, ranging from 10.53 ± 0.78 Ma to 14.4 ± 1.1 Ma, with no clear systematic variation across the section (Table 3).

These cooling ages are distinctly younger than the emplacement age of the Kuday AEG granite (~28 Ma) and that of the AEG group in general (~28-23 Ma). In contrast, these ages are comparable to the emplacement ages of the DPP granites (~14.5 - 9 Ma), with Ar-Ar single and multi-grain biotite analysis for the Kouwu and Mabja DPPs also revealing biotite cooling ages of 8.8 ± 0.3 Ma to 8.0 ± 0.5 Ma, coeval with the U-Pb crystallisation age of the Mabja granite, the youngest of the DPPs (Table 2). These observations are consistent with the data of both the transect of the Sakya¹ dome immediately south west of the Kuday area, with mica closure ages of 12.85 ± 0.13 Ma to 17.0 ± 0.19 Ma (Lee et al., 2006), with a transect across the Kampa

dome immediately to the east, with mica Ar closure ages ~14.6 Ma (Quigley et al., 2006), and for the uplift history of the Leo Pargil dome of the Western Himalaya with mica Ar closure ages of 14-16 Ma (Thiede et al., 2006)

DISCUSSION

Geological significance of U-Pb ages

Composite grains of accessory minerals with older cores and younger rims are common in many North Himalayan granites. A thin rim on the dated phase may result from metamorphic recrystallisation rather than representing a fresh igneous crystallisation event; in this case the rim represents metamorphic zircon growth and the upper intercept is commonly interpreted as the crystallisation age. For the North Himalayan granites, this interpretation might suggest that upper intercepts represent Lower Paleozoic (Cambrian) melts that have undergone Tertiary deformation and heating. We reject this interpretation on the following grounds: (i) thermal diffusion rates are insufficient to grow a detectable rim during metamorphism at the relevant temperatures of $< 770\text{ }^{\circ}\text{C}$ (Watson and Harrison, 1983); (ii) field relations indicate that the younger granites cross-cut the Tethyan sedimentary sequence and thus cannot be Paleozoic in age; (iii) Cambrian augengneisses of the dome cores (such as the Kangmar gneiss) are not observed to cross-cut the Tethyan metasediments; (iv) the Cambrian gneiss is characterised by distinct geochemical trends compared to the younger leucogranites (Zhang et al., 2004). In general, it is unlikely that zircon overgrowths represent metamorphism rather than crystallisation in crustally-derived granites, since at their low temperatures of formation only a small fraction of source grains need be resorbed into the melt before Zr saturation is reached (Watson and

Harrison, 1983). As a result, the majority of ‘inherited’ zircon grains are likely to be preserved intact, resulting in grains with large, often euhedral, inherited cores and narrow overgrowths.

We conclude that the U-Pb Tertiary intercept ages (Table 2) are representative of melt crystallisation events rather than of metamorphic resetting, and interpret discordant zircon analyses as representing mixed ages resulting from either laser spatter or by transgression of the ablation raster across growth zones. Thus the upper-intercept ages from 573 ± 24 Ma to 473 ± 17 Ma are inheritance ages, indicative of an early Paleozoic source. The presence of Neoproterozoic ages in at least two AEG indicates yet older inherited detrital material is present in the source. The lower-intercept ages define two groupings of crystallisation ages; an Oligocene age span for the AEG granites, ranging from ca. 28.1 ± 0.4 Ma (Kuday) to 22.6 ± 0.4 Ma (Kua), and a mid-late Miocene age span for the DPP granites, ranging from 14.4 ± 0.6 Ma (Gomdre) to 9.8 ± 0.2 Ma (Mabja).

Conditions of melt formation

Major elements of the analysed leucogranites suggest that the DPP are strongly to mildly peraluminous whereas AEG compositions are mildly peraluminous to metaluminous. The high-Ca compositions that distinguish AEG granite compositions suggest a preference for plagioclase over muscovite in the muscovite melt reaction:

which is favoured by increasing pressure (Patiño Douce and Harris, 1998) indicating that the protoliths of the AEG were deeper than those of the DPP at the time of their melting. Higher pressures (>8 Kbar) are supported by the rare presence of kyanite in the Kuday granite (Zhang et al., 2004).

The source of both groups of granites lies in the High Himalayan crystallines of the Indian Plate (Fig. 10b). However, whereas the DPP plutons must be sourced in the metasedimentary formations, the AEG field maps closely in Sr-Nd isotope space onto the field of Cambrian granitic gneisses, as represented in the Himalaya of NW India (Miller et al. 2001) and the gneisses of the Kangmar dome (Fig. 1). The presence of inherited zircons of 900-1100 Ma from both the Kuday granite and the Cambrian granite gneisses of NW India (Miller et al., 2001) supports this inference. Quigley et al. (2006) also speculated that the high volume of Tertiary intrusions exposed within the Sakya dome relative to those domes to the east (Watts et al., 2005), implies that the Sakya dome exposes deeper crustal levels. This may have implications for the geometry of the Indian crust prior to plate collision, or alternatively suggests localised mid crustal ramping during the Oligocene.

The observation that AEGs are distinguished from DPPs by low LREE/HREE ratios and high Y (Fig. 9) suggests either that garnet crystallized in the AEG partial melts such that HREE and Y from the protolith were strongly partitioned into the garnet structure or that xenocrystic garnets have been entrained from source metasediments. However, garnets from AEG assemblages are spessartine-rich; high spessartine garnets are generally characteristic of an igneous origin (Harris et al., 1992). In contrast, garnets from the metapelitic assemblages within the dome are depleted in both spessartine and grossular, and are thus typical of metapelitic

assemblages. We conclude that the garnets from the granites are not xenocrystic, at least not from a metasedimentary source. They most likely represent a product of the melt reaction, requiring biotite to be a reactant. The biotite breakdown reaction

has been experimentally determined for metapelitic lithologies from the High Himalayan Series at temperatures above 700 °C (consistent with the low temperatures obtained by zircon thermometry) and pressures of *ca.*10 kbars, during fluid fluxing (Patiño Douce and Harris, 1998). Fluid fluxing is consistent with the anastomosing structure of the granites (wet melts cannot rise far through the crust without crystallising), and also with myrmekitic intergrowth, characteristic of the AEGs in thin section, since myrmekite growth is promoted by the presence of water at grain boundaries (Hibbard, 1979, I. Parsons *pers.comms.* June 2007). Fluid fluxing is also consistent with the Rb, Sr, Ba systematics since Rb/Sr ratios are relatively invariant during melt formation at higher fluid-phase activities, as discussed above. Moreover, since the reaction requires biotite, rather than muscovite, in the source it allows the possibility that these granites were sourced from melting of the Cambrian granitic gneisses (which are largely muscovite absent), as indicated by the Sr-Nd data (Fig. 10b).

Structural observations on the Sakya dome suggest the following deformational history. Pervasive early top-to-south shear under prograde metamorphic conditions associated with crustal thickening was accompanied by emplacement of the AEG granites. Subsequent top-to-north shear was partitioned at lower grade mainly into the

weaker Tethyan Sedimentary Series cover rocks, and may equally be representative of the initiation of extrusion of a melt-weakened crustal channel or emplacement of a mid-crustal thrust wedge. Later top-to-south shear features have so far only been observed on the south-west fringe of the Sakya dome, and may represent deformation localized to DPP margins. The Sakya dome structures were finally transected by mainly N-S trending, brittle normal faults. The alternating ductile shear sense observed at the margins of the Sakya dome could be explained either by tectonic wedge models (e.g. Price, 1986; Webb et al., 2007), or as the upper margin of a ductile channel (Beaumont et al., 2001). Crucially, the AEGs were emplaced before top-to-north shearing along the dome core/Tethyan sediment contact. Therefore this deformation phase (either onset of channel flow or tectonic wedge advance) must be younger than ~28.1 Ma (at Kuday) and ~22.6 Ma (at Kua).

The formation of the AEG melts is consistent with the initial melt weakening required to facilitate ductile flow and the southward transport of mid-crustal material (Beaumont et al., 2004) as recent steady-state fluid-mechanical studies suggest that effective viscosities can be reduced by about 50% by low-degree partial melting, with a melt fraction of ca. 0.4 (Holtzman et al. 2005). Even at much lower melt fractions ($F=0.07$) a dramatic loss in aggregate strength is observed (Rosenberg & Handy 2005). This compares well with the observed velocity structure within the low-velocity zone beneath southern Tibet that is consistent with a melt fraction *ca.* 0.07 to 0.12 (Yang et al. 2003) and with magnetotelluric data which suggest a melt fraction of 0.05 to 0.14 to account for the low-resistivity layer (Unsworth et al. 2005).

Although Lee and Whitehouse (2007) have derived a flow onset age of 35 Ma, this inference is based on a dated leucosome taken from an isolated migmatite that is

unconnected to the D₃ top-to-north deformation event. In contrast we present intrusion ages for the AEGs (28–23 Ma) that immediately predate shear-sense reversal at this time. Although it remains probable within a heterogeneous crust that some lithologies had attained temperatures and pressures sufficient for anatexis by ~35Ma (or 39 Ma; see Prince et al, 2001), we suggest that insufficient melt volume had formed to lower the viscosity to the required threshold for crustal flow until after 28 Ma, as this process requires a regionally distributed zone of melting in order to lower the crustal viscosity on a length-scale of decimeters, if not kilometers. During tectonic thickening of the mid to lower crustal pile, be it by lower crustal underplating or by tectonic wedging, self-heating in response to crustal thickening would promote melting, particularly under vapour-present conditions, as inferred from AEG trace-element compositions, texture and structural setting. Source heterogeneity and localized fluid infiltration could give rise to the observed spread in AEG emplacement ages (28-23 Ma), and also to the occurrence of migmatite formation as early as 35 Ma.

In contrast to the AEGs, the DPP intrusives were emplaced as discrete plutons under more brittle conditions at higher crustal levels, forming limited thermal aureoles (Lee et al., 2004) and late, localized top-to-south buoyancy-related structures as seen at the margins of the Gomdre pluton. Their elevated Rb/Sr ratios at low Ba abundances, coupled with low melt temperatures, indicate small-melt-fraction anatexis during vapour-absent muscovite breakdown by the reaction:

at temperatures of 750 - 770°C and pressures of 6 - 8 kbar, as described by Patiño Douce and Harris (1998) for the Miocene leucogranites of the High Himalaya. The marked negative slope of this reaction in pressure-temperature space not only allows melts to rise significant distances through the crust before crystallizing but also enables anatexis through decompression melting. The source requires the widespread presence of muscovite as in the metasedimentary schists of the High Himalaya, a source isotopically consistent with Sr-Nd data from the DPPs (Fig. 10b). The elevated $^{87}\text{Sr}/^{86}\text{Sr}$ and low $\epsilon_{\text{Nd}}(10)$ of the Mabja granite suggests a source from migmatites found near the top of the Indian slab in the central Himalaya (Inger and Harris, 1993).

Dome formation and thermal structure

The Ar-Ar cooling ages obtained in this study suggest little if any spatial variation across the traverse. Lee et al. (2006) observed that Ar-Ar ages from the western margin of the Sakya dome increased down-section from 13 Ma to 17 Ma and then decreased to 13 Ma at the lowest structural levels, into which the Donggong granite had been emplaced. In this study we have reported a U-Pb emplacement age of 14.5 ± 0.9 Ma for Donggong samples, coeval with the younger Ar-Ar ages of 13.54 ± 0.06 Ma and 13.48 ± 0.12 Ma for muscovite and biotite respectively at the deepest structural levels (Lee et al., 2006). This would suggest that the apparent younging is due to reheating of the sampled rocks on emplacement of the Donggong pluton. If overprinting by younger plutons is taken into account, some younging towards the margins of the dome may be inferred from the data of Lee et al. (2006). A study from the adjacent Kampa granite (Quigley et al., 2006) also found the youngest biotite ages of *ca.* 13.7 Ma immediately below the granite-metasediment contact. However,

cooling age trends across published traverses vary by only *ca* 1.5 million years, which may arise from either Neogene faulting and/or overprinting by granite intrusions, both of which are abundantly evident. In the absence of evidence for distinct younging of cooling ages toward the centre of the dome, as would be expected if doming were a response to an extruding channel, or of significant younging toward the dome fringes, as may be the case if uplift were entirely buoyancy driven, we conclude that the timing of crustal uplift cannot be unequivocally linked to the domal geometry of the Sakya dome as a whole.

For the Oligocene (AEG) bodies the Ar-Ar cooling ages are significantly younger than emplacement ages established by this study (28–23 Ma). This implies that (i) AEG emplacement occurred at a depth >10 km (assuming an Ar closure temperature of 350 °C for biotite, and a transient geotherm in tectonically thickened crust (Jamieson et al., 2004)); and (ii) unroofing/exhumation of this complex to depths < 10 km occurred a significant time interval (> 8 million years) after AEG emplacement. In addition, peak metamorphic grade mineral assemblages associated with the main foliation (S_2) are contemporaneous with AEG emplacement. Because S_2 foliation defining an overall domal geometry is recorded by early AEGs but not by younger, unfoliated AEGs and the margins between the two facies are diffuse, both AEG facies must represent subsets of the same magmatic suite that have experienced different degrees of sub solidus strain at time of emplacement. This suggests that at least some doming occurred at depth while emplacement of AEG melts was still ongoing (Fig. 13a) – a scenario compatible with emplacement and doming during crustal underplating of a High Himalayan Series protolith wedge (Yin, 2006). Exhumation and retrograde metamorphism occurred syn- to post- D_3 deformation (e.g. ductile shear

bands), with later D₄ and D₅ ductile to brittle behaviour indicative of exhumation to shallower crustal levels with time. Ar-Ar single and multi-grain biotite analysis for the individual DPPs reveal biotite cooling ages range from 8.8 to 8.0 Ma (Table 2) which overlap with both the U-Pb emplacement ages of the DPP granites (14.5 to 8.8 Ma) and hence the localized D₄ deformation potentially associated with DPP emplacement. The metamorphic aureoles of such intrusives suggest they were emplaced into significantly cooler crust (Lee et al., 2004). We infer that, as DPP granites cross-cut AEG-related domal structures, such as those seen at Kuday, presently exposed crust had acquired at least some degree of domal geometry prior to cooling through the biotite closure temperature (*ca.* 350 °C) at 14.4 to 8 Ma.

AEG-related doming structures may well initially be a response to lower crustal underplating, duplexing or tectonic wedging (Yin, 2006) of Indian metasedimentary crustal successions (such as the Lesser Himalayan Series and High Himalayan Series) beneath the Tethyan metasedimentary wedge, with the Asian crustal block as the backstop or over-riding block. However, doming, and hence decompression of the upper crust, is also envisaged by some formulations of the channel-flow model of Beaumont et al. (2004) in response to the detachment and extension of weak upper crust on the hinterland side of the erosion front leading to doming of the channel beneath the extending and thinning upper crust. If doming results in ‘freezing’ of a former extruding channel as temperatures drop below *ca.* 700 °C, buckling of the former channel will provide an additional mechanism for local Miocene uplift, with continued crustal extrusion at deeper structural levels not precluded. Nonetheless, one would not expect such consistency in dome exhumation ages across the North Himalayan antiformal structure if individual dome uplift was due to buckling of

discrete frozen remnant channels. Regional scale ramping may, on the other hand, be a consequence of large scale tectonic wedge emplacement, consistent not only with uplift of the domes at *ca.* 15 Ma and rapid exhumation documented between 16 -14 Ma in the Leo Pargil dome (Thiede et al., 2006), but also with observations of alternating shear sense on the South Tibetan Detachment and mid Miocene movement recorded on the Great Counter Thrust (Yin 2006, Webb et al., 2007). Equally the uplift of a regional scale antiform (e.g. the North Himalayan antiformal structure), may be a plate flexural response to slab steepening, override and/or subsequent break-off of the subducted Indian lithosphere, following AEG emplacement, an interpretation consistent with both tomographic observations (Replumaz and Tapponier, 2003), and with the formation and emplacement, beginning *ca.* 19 Ma, of shoshonitic melts derived from the sub-continental mantle beneath southern Tibet (Maheo et al., 2002, Williams et al., 2001; 2004). A continuation of this process would account for the emplacement of anatectic dikes of Asian plate affinity from 17 – 9 Ma (Fig. 13b), younging from north to south across the suture zone into the Tethyan sediments of the Sakya area (Williams et al., 2004; King et al., 2007). Although flexure of the lithosphere has been postulated as relating to isostatic rebound due to slab break-off (e.g. Maheo et al., 2002; Mugnier and Hyughe, 2006), the current lack of evidence for asthenospheric melts that would be expected if slab break-off were involved, would suggest that only limited lithospheric mantle thinning occurred at this time (King et al., 2007).

Structures relating to early AEG emplacement suggest initial doming, plausibly due to tectonic underplating of Indian mid-lower crust, prior to top-to-north shear sense crustal deformation in the hinterland Tethyan Sedimentary Series, and certainly

prior to the earliest formation of DPP melts at *ca.* 15 Ma, which coincides with a period of cooling, orogen-parallel extension and rapid exhumation (Thiede et al., 2006). Whereas the North Himalayan antiform is a first-order regional structure, probably related to flexural bending due to slab steepening, the individual domes are second-order features imposed on that structure by localized faulting, and/or pre-existing sub-surface domal structures.

We propose that following initial doming, further uplift of Indian mid-crustal rocks, possibly related to slab steepening, triggered decompression melting at 15-9 Ma, with subsequent ascent and emplacement of the DPP granitic melts into the upper crust, imposing additional buoyancy-related structural overprints on areas proximal to DPP intrusions. Over a similar period (12-9 Ma) suites of north-south trending dacite dikes were emplaced south of the suture (Fig. 13b), some of which are associated with north-south graben (King et al., 2007). This period of east-west upper crustal extension culminated with brittle displacement along the north-south striking normal faults in the Sakya dome (Fig. 2), and appears to be a regional phenomenon, since coeval east-west extension and associated dacitic to shoshonitic magmatism are recorded north of the Indus-Yarlung-Tsangpo suture zone far into central Tibet (Turner et al, 1996; Miller et al., 2001; Williams et al, 2001, 2004).

CONCLUSIONS

This study of the granites of the Sakya dome has clearly identified two groups of granitic intrusions that have distinct field relations, geochemistry, geochronology and tectonic implications. The oldest group of granites (AEG) were emplaced between 28 and 23 Ma. They have irregular, anastomosing contacts and were generated by

anatexis of the granitic gneiss that now forms the core of several of the North Himalayan gneiss domes. The compositions of AEGs are characterised by high HREE/LREE ratios and Y abundances, which can be related to the presence of biotite in the melt reaction under conditions of high fluid-phase activity, and the consequent formation of melt and peritectic garnet. The melts aggregated to form the observed anastomosing bodies, but did not rise significantly through the crust. During this period, underplating and thickening of the mid crust of the Indian Plate resulted in fluid-present partial melting where local fluid influx and the presence of fusible lithologies permitted. Such a reduction in bulk viscosity could have enhanced, or initiated, flow of a mid-crustal channel (Fig. 13a). A reversal in shear sense along the dome core/Tethyan metasediment boundary postdates the onset of AEG anatexis, and is consistent with the southwards extrusion of a low viscosity channel below this upper bounding shear zone. However, this is not necessarily a unique solution, as AEG-related domal structures could also reflect underplating or tectonic wedge emplacement, with secondary uplift in response to plate flexure.

The younger group of granites (DPP) form discrete plutons that have been emplaced between 15 and 9 Ma, from metasedimentary sources also within the Indian Plate. Geochemically these granites are distinguished by their high Rb/Sr ratios at low Ba abundances, indicative of fluid-absent melting, and suggesting decompression as a cause of anatexis.

We suggest that the timing and mode of origin of the Miocene DPP intrusions are more consistent with localized, buoyancy-driven emplacement of decompression melts than with regional-scale crustal flow processes. Conversely, AEG anatectic conditions were optimal for the initiation of regional crustal flow during the

Oligocene. We further propose that whilst doming at depth is a possible consequence of channel flow, it is likely to have been enhanced during the mid-Miocene by the plate flexural response to slab steepening (Fig. 13b) which also provides a mechanism for rapid post-dome formation uplift, late movement on the Great Counter Thrust and decompression between 15 and 9 Ma, culminating in E-W crustal extension as evidenced by the north-trending graben and associated mid-Miocene dike swarms intruding both sides of the suture in southern Tibet.

APPENDIX

GEOCHEMICAL METHODOLOGY

Elemental analysis

Whole-rock major and trace element, and Sr-Nd isotope analyses were undertaken to shed light on the provenance of the granitic melts. We used LA-PIMMS analysis of monazite, xenotime and zircon and laser spot Ar-Ar analysis of biotite to ascertain granite intrusion and cooling ages respectively. Whole-rock major and trace elements were analysed on an ARL Fisons wavelength-dispersive XRF spectrometer at the Open University. Major elements were determined using glass discs prepared by fusing powdered samples with Spectroflux 105. Rb, Sr, Ba, Zr, Y and Nb were determined from pressed powder pellets. Additional trace elements (REE, Ta, Th and U) were analysed by ICP-MS. 100 mg aliquots of powdered samples and standards were dissolved in PFA pressure vessels using 3ml TD HF and 1ml TD HNO₃ for a total of 48 hours on a hotplate at temperature of 130 °C. Solutions were sonicated for a period of 20 minutes, checked for clarity and evaporated to incipient dryness before

being redissolved in 2 ml concentrated TD HNO₃ and 4 ml MilliQ deionised water at 130 °C for 24 hours (plus a 20 minute sonication). This step was repeated with a further 24 hour 3 ml concentrated TD HNO₃ – 4 ml MilliQ deionised water dissolution. After further evaporation to incipient dryness, samples were dissolved in 1 ml TD HNO₃ and diluted to 100 ml using 18 MΩ deionised water, and stored in clean polypropylene bottles. The solutions were analysed using an Agilent 7500s ICP-MS instrument, fitted with a Babington nebuliser operating at a flow rate of 0.4 ml min⁻¹. Internal standards (Be, Rh, In, Tm, Re and Bi) were added on-line using a second peristaltic pump and residual drift for individual elements was assessed and corrected using repeat analyses of selected sample solutions. Standards used for calibration were BHVO-1, AC2/10, AGV-1, DNC-1, W-2, G-2 and RGM-1 as calibrated by Eggins et al. (1997) and within run precision was determined using repeat analyses of AC2/10 and G-2. In general, within-run precision was better than 2% r.s.d. for all elements and often better than 1%.

Sr-Nd isotopic analysis

Nd and Sr isotope ratios were analysed on a ThermoFinnigan Triton thermal ionization mass spectrometer (TIMS) at the Department of Earth Sciences, Open University. Standard sample preparation and isotopic analytical techniques are described in Cohen et al. (1988) and Charlier et al. (2006). Strontium was loaded in phosphoric acid on single Ta filaments, and the measured ⁸⁷Sr/⁸⁶Sr ratios were corrected for instrumental mass fractionation using ⁸⁷Sr/⁸⁸Sr = 0.1194 and the exponential fractionation law. Repeat analyses of the NBS987 Sr standard gave ⁸⁷Sr/⁸⁶Sr ratios of 0.710243 ± 0.000014 (2σ) over the analysis period. Total

procedural Sr blank was 20pg. $^{87}\text{Rb}/^{86}\text{Sr}$ ratios were calculated from elemental ratios obtained by XRF. Neodymium was loaded on double Re filaments and run as metal ions. $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were normalised to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. Repeat analyses of the Johnson-Matthey internal standard gave 0.511817 ± 0.000006 (2σ) for the period of sample analysis, a value for BHVO-2 (external standard) of 0.512976 ± 0.000014 and for LaJolla of 0.511845 ± 0.000005 . Total procedural Nd blanks were <1 ng. $^{147}\text{Sm}/^{144}\text{Nd}$ ratios were calculated from elemental ratios obtained from ICP-MS.

Ar isotope analysis

For ^{40}Ar - ^{39}Ar isotope analysis, single biotite mineral grains were analysed using the Infrared laserprobe single-grain fusion technique. Although the granite samples contained both muscovite and biotite, muscovite grains proved unsuitable for analysis being either too small or impure. Thirty biotite grains of grain size ca. 0.5 to 1mm were separated from each sample by standard mechanical crushing and sieving techniques followed by final hand-picking using tweezers and a binocular microscope to ensure purity. Altered or discoloured grains were discounted. Each biotite separate was cleaned ultrasonically in alternate ethanol and de-ionised water, prior to packaging in aluminium foil. Each individual foil packet contained a biotite separate, and was bracketed by GA1550 biotite standards to monitor neutron flux during irradiation.

The samples were irradiated for 25 hours at the McMaster Reactor (Canada) in position 5C with cadmium shielding. Neutron flux was monitored using biotite mineral standard GA1550 which has an age of 98.8 ± 0.5 Ma (Renne et al. 1998). The resulting J value was 0.00639 ± 0.000032 . Results were corrected for ^{37}Ar decay, and

neutron-induced interference reactions using the following correction factors: $(^{39}\text{Ar}/^{37}\text{Ar})\text{Ca} = 0.00065$, $(^{36}\text{Ar}/^{37}\text{Ar})\text{Ca} = 0.000264$, and $(^{40}\text{Ar}/^{39}\text{Ar})\text{K} = 0.0085$, based on analyses of Ca and K salts.

The irradiated samples were loaded into an ultra-high vacuum system and mounted on a New Wave Research UP-213 stage. A 1064nm CW Nd-YAG Spectron laser was focused into the sample chamber and was used to fuse individual biotite grains such that each grain yielded a single age, referred to as a single grain age. For samples with smaller grains, several grains were fused to ensure the release of measurable gas volumes. These analyses are referred to as multi-grain analyses. After each analysis the extracted gases were cleaned for 5 minutes using two SAES AP-10 getters running at 450° C and room temperature prior to automated inlet into an MAP 215-50 noble gas mass spectrometer. System blanks were measured before and after every two sample analyses and the mass discrimination value for atmospheric $^{40}\text{Ar}/^{36}\text{Ar}$ was measured at 283. Isotopes of ^{40}Ar , ^{39}Ar , ^{38}Ar , ^{37}Ar , and ^{36}Ar were measured fifteen times per scan, and for ten scans, the final measurements are extrapolations back to the inlet time. All ages are reported at the 2σ level and include a 0.5% error on the J value; weighted mean ages are calculated using Isoplot 3a (Ludwig, 2003).

U-Pb isotope analysis

An *in situ* U-Th-Pb dating methodology was essential for this study given the abundance of inherited zircon with Paleozoic cores in leucogranites of the Himalayan orogen. In order to determine crystallisation ages, the method of Horstwood et al. (2003) was used using laser rastering with shallow pit aspect ratio, using laser ablation microsampling and multiple collector ICP-MS instrumentation.

Accessory minerals were separated from their host rock using conventional techniques (crushing, milling, sieving, Rogers® table, heavy liquid separation (Methylene Iodide), and Frantz magnetic separation). Mineral grains were then hand-picked under alcohol; in the case of zircon, morphologies were selected based on their probability of either being pristine magmatic zircons formed wholly in the latest/youngest melt generation event experienced by the rock, or grains with a demonstrably inherited core and a clearly overgrown 'rim' (possibly from the latest melt generation event). Their morphologies ranged from very slim pristine euhedral needles with aspect ratios >8:1 (morphology N) to squat diamond-shaped grains with spherical to subhedral, clearly inherited, cores and sharp clear euhedral terminations. Two methods of grain mounting were used to optimise the grain aspect for successful analysis.

Conventional grain mounts

For zircon morphologies where it was thought that adequate target sample area was achievable in cross-section, and for all other accessory phases of interest (xenotime, monazite and allanite), a conventional approach to grain mounting was taken in which grains embedded in epoxy resin mounts were polished to expose an internal surface and rinsed with 2% HNO₃ prior to analysis. Zircons from these mounts were imaged at electron microscopy facilities of the Open University (BSE) and Leicester University (for CL).

Flat grain mounts

Where igneous growth zones <25µm were suspected to be surrounding an older core an alternative strategy was used. Using relatively large euhedral grains with well-

developed crystal faces, grains were mounted directly onto double-sided sticky-tape attached to the surface of a resin disc, ensuring that the flattest surface possible was available for laser ablation. A diffused laser spot was used to ablate a line or box raster (~50-200 μm length) across the external surface, enabling thin zircon rim material to be milled to a depth of $\sim < 5 \mu\text{m}$ for analysis, minimizing the ablation of the underlying inherited core. This method also has the advantage of avoiding inter-element fractionation with increasing crater depth and zonal cross-contamination from material spatter that results from analysing stationary spots.

Laser sampling and mass spectrometry

Over the period June 2004 to October 2006 two mass spectrometer set-ups were used. The early data were acquired using a VG(Thermo) Elemental Axiom double-focussing MC-ICP-MS coupled to a New Wave Research Microprobe II, 266 nm Nd:YAG laser ablation system. A Cetac Technologies Aridus nebuliser was used to simultaneously introduce a $\text{Th}-^{235}\text{U}$ solution to the plasma to allow correction for instrumental mass bias and monitoring of inter-sample variation of plasma induced inter-elemental fractionation. The Axiom was fitted with a multiple-Faraday (F) main collector array and when using peak switching, all isotopes from 200 to 208 and 235 and 238 were measured, with 204 being measured in an axial electron multiplier in a manner identical to that described by Horstwood et al. (2003). Xenotime samples were run with laser spot sizes of 72 μm at 5 Hz, 60-80% power. Zircon samples typically ran with spot size 35 μm at 10 Hz, 80% power.

Data from August 2006 onwards were acquired using a NuPlasma MC-ICP-MS with a New Wave 193 nm solid state laser and Nu Instruments nebuliser set-up (with

Tl- ^{235}U solution) similar to that described above. The collector configuration was static with multiple ion counting and faraday cups. The detectors consisted of ^{202}Hg (F-L5); ^{203}Tl (F-L4); ^{204}Pb (EM-IC2); ^{205}Tl (F-L3); ^{206}Pb (EM-IC1); ^{207}Pb (EM-IC0); ^{235}U (F-H6) and ^{238}U (F-exH). Typical laser spot size was 25 μm , at 5 Hz and power settings were between 30-55% for samples and 50-65% for standards.

Data were normalized to one or more external ablation standards (zircon 91500; Manangotry monazite) to correct for mass and elemental fractionation during ablation, and machine drift. Concordia were calculated using Isoplot version 3 beta (Ludwig 2003). Decay constants used for ^{238}U ; ^{235}U and ^{232}Th are those recommended by Steiger and Jäger (1977).

During analysis sessions on the Axiom and Nu Plasma, standards were analysed in a ratio of approximately 3:1 (samples:standards). Samples were analysed in blocks of time between gas tuning optimisation, and normalised to the mean value of standards run during these blocks. These ‘blocks’ varied between 1 – 8 hours depending on the reproducibility of the standards in terms of $^{206}\text{Pb}/^{238}\text{U}$. The standard deviation of $^{206}\text{Pb}/^{238}\text{U}$ on standards during blocks varied from as little as 0.7% to about 2%, and this ‘external reproducibility’ factor was quadratically incorporated into the final uncertainty estimation for all individual analyses. Final isotope ratios used in plotting on Concordia diagrams incorporated a correction for common Pb when the ^{204}Pb intensity (after subtraction of the ^{204}Hg signal) was significant. In some of the analyses using the more sensitive Nu Plasma (permitting multiple ion counting), the amount of ablated samples was as little as 10 nanograms. Given the instrumental background from ^{204}Hg , it was difficult to measure the ^{204}Pb contribution from common Pb accurately in such small samples. Hence a Tera-Wasserberg diagram

was used on multiple analyses to define the intercept age independent of ^{204}Pb abundances (see figures for Donggong (11a), Gomdre (11b), Kua xenotime (9c)).

Inspection of the data indicates that many spots on zircon produced discordant results defining an array connecting an upper and lower intercept. This style of data is a consequence of ablation of variable amounts of two-age components during the spot or raster analysis. In particular when a raster approach of sampling is used, there is marked variability within the analysis of the proportions of the two contrasting age components, a situation that commonly results in a mean standard deviation in $^{206}\text{Pb}/^{238}\text{U}$ of the spots much larger (i.e. >2%) than is seen in a homogeneous standard (<1%). Such a data spread is an indication of heterogeneity with each analysis.

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FIGURE CAPTIONS

Figure 1 Geological sketch map of southern Tibet showing North Himalayan gneiss domes. ITSZ = Indus-Yarlung-Tsangpo suture zone, STDS = South Tibetan detachment system, GCT – Great Counter Thrust. Box indicates location of Fig. 2.

Figure 2 Geological map of the Sakya Dome. $\oplus$ = granite intrusion: solid line delineates margins of DPPs; anastomosing nature of AEGs indicated by absence of bounding line, $\times$ = pegmatite swarm.

Figure 3 Caption: Locality key of lithological and structural observations described in upcoming figures 4 through 7. Lithological ornament as for Figure 2.

Figure 4 Contact relationships between Tethyan Sedimentary Series and Kuday granite (AEG) from northern Sakya Dome, seen from east side of the Kuday gorge. (a) AEGs intruding orthogneiss, view looking north from sample site JK3/05; and (b) the same relationship in detail with annotated sketch showing the complex timing relationships between AEG, foliated AEG, Kuday orthogneiss, restite, and AEG pegmatites 1 and 2, from same sample site JK3/05; (c) view looking west, north of contact. (d) with overlay showing granite (AEG) dike distribution (indicated by crosses). The Miocene dacitic dikes (King et al., 2007) are indicated by irregular black lines. Distance from contact of lower granite to top of section is 50 m.

Figure 5 Tectonic indicators for D2 during Kuday (AEG) intrusion into the Tethyan Sedimentary Series cover. (a) Sheared AEG pegmatite vein, view looking east; darker wedge is soil cover. (b) with overlay showing kinematics of pegmatite deformation (top-to-south); lower heavy boundary marks contact between core lithologies (~5m thick AEG sheet) and garnet mica schists of the metasedimentary cover (TSS); upper boundary marks intrusive contact between TSS and AEG dike. (c) main image shows view along the main contact (see inset) towards the north-west, and depicts the rotational kinematics (top-to-south) of AEG dike deformation assuming vertical intrusion. (d) mylonitic metasediments, marbles and amphibolites (prominent foreground exposure) ~200m above the domal Kuday gneiss-granite complex/TSS contact (surface outlined by dashed line). View towards west from the eastern side of the Kuday gorge.

Figure 6 Top-to-north shear band development in the Tethyan Sedimentary Series above the contact with the Kuday granite (AEG) from northern Sakya Dome, (a) view towards west on the west side of the Kuday gorge with (b) annotated overlay. S_2 (white dashed ornament) is main foliation; black dashed lines indicate C-surfaces of S_3 shear bands. (c) view towards east on the east side of the Kuday gorge with (d) annotated overlay (ornaments as in part (b)). (e) Plane polarised light and (f) Cross-polarised photomicrograph of top-to-north shear bands (S_3) deflecting main foliation (S_2) at microstructural scale in pre- D_3 -deformational (M_1) garnet mica schists. Garnet porphyroblast size 1 – 5mm.

Figure 7 Photomicrographs of Tethyan metasedimentary samples show microstructural relationships between metamorphic porphyroblasts and foliation. (a) pre- to syn-S₂ deformational chloritoid and (b) garnet with top-to-north rotation from intercalated schists of the chloritoid zone. (c) pre-S₂ deformational garnet with syn-S₂ staurolite after chloritoid. (d) Retrograde chlorite filled pre-S₂ garnet pressure shadows in garnet mica schist

Figure 8 Chondrite-normalised plot for rare-earth-element compositions for granites intruding the Sakya dome. Dashed lines indicate AEG, solid lines DPP. Data from Table 1 and Zhang et al. (2004).

Figure 9 La/Yb vs Y plot of Sakya dome granites. Anastomosing, equigranular granites (AEG) = ■, Discrete, porphyritic plutons (DPP) = ○. Data from Table 1 and Zhang et al. (2004).

Figure 10 Sr-Nd isotopic ratios calculated at 10 Ma. (a) Potential melt source fields from published data. Asian plate (Lhasa terrane gneisses and eclogites, Gangdese plutons, and tertiary volcanic and intrusives = ◇; Tethyan Sedimentary Series (TSS) = ■; Indian crust metasedimentary formations (High Himalaya = ▲; Lesser Himalaya = △); CGG (✕) indicates Cambrian granitic gneiss intruding the High Himalaya formations. Published data from Ahmad et al., 2000 and references therein; Williams et al., 2001; Zhang et al. 2004; Richards et al., 2005; Richards et al., 2006; Mo et al., 2007; Li et al., 2009; Zhu et al., 2009a; Zhu et al., 2009b; Zhu et al.,

2009c; Zhang et al. (in press) (b) North Himalayan granites; AEG = ○, DPP = ●.
Data from Table 1 and Zhang et al. (2004).

Figure 11 Concordia diagrams for dated accessory phases for AEG bodies; (a) Kunday, (b) Wing, (c) Kua, (d) Lijun. CL and SEM images show accessory mineral grains yielding concordant analyses, numbered boxes and circles represent the analysis number (i.e. JK3/05z1-2 in Table DR1) and box-raster and spot dimensions respectively. Z = zircon; m = monazite; x = xenotime. Errors depicted at 1σ .

Figure 12 Concordia diagrams for dated accessory phases for DPP bodies; (a) Donggong, (b) Gomdre, (c) Mabja. CL and digital photomicrograph images show accessory mineral grains yielding concordant analyses, numbered boxes and circles represent the analysis number (i.e. JK4/12az1-1 in Table DR1) and box-raster and spot dimensions respectively. Errors depicted at 1σ .

Figure 13 Cartoon showing proposed tectonic evolution of southern Tibet (a) from 28-23 Ma; AEG developed within partial melt zone of Indian mid crust; doming and cooling of AEG top-to-north structures may be attributed to relative southwards motion of a melt-weakened crustal channel or equally in response to plate flexure on steepening of the subducted Indian slab (Replumaz et al., 2004); (b) from 15-9 Ma during dome exhumation, decompression and formation of DPP. SCLM = sub-continental lithospheric mantle, MCT = Main Central thrust, STDS = South Tibetan detachment system; GCT – Great Counter Thrust. Vertical bars indicate dacite dike emplacement (King et al., 2007).

TABLE 1: Major, trace and Sr, Nd isotopic data for North Himalayan Granites
AEG Granites

Sample Age (Ma)	Data from Zhang et al 2004				Data from JAK 2003				Data from JAK 2004							
	Kuday		Kuday		Kuday		Kuday		Kuday		Kuday		Lijun		Kua	
	T100	T101	T104	T105	JK3/05	JK3/08	JK3/12a	JK3/13b	TO3/29x	TO3/31x	TO3/33x	JK4/07a	JK4/07a	JK4/08	JK4/08a	Kua
	AEG	AEG	AEG	AEG	AEG	AEG	AEG	AEG	AEG + tur	AEG	AEG	AEG	AEG	AEG	AEG	AEG
SiO ₂	74.04	73.65	73.90	73.94	70.95	74.69	73.14	74.05	77.44	76.97	77.72	76.16	67.28	76.20	63.99	63.99
TiO ₂	0.11	0.01	0.12	0.07	0.28	0.06	0.11	0.04	0.12	0.07	0.13	0.13	0.69	0.05	0.57	0.57
Al ₂ O ₃	14.78	14.92	14.70	15.11	14.36	14.99	14.89	15.19	12.74	13.27	12.57	12.63	15.33	14.84	15.86	15.86
Fe ₂ O ₃	0.04	0.13	0.19	0.10	1.93	0.39	0.98	0.77	1.02	0.25	1.00	1.68	5.15	0.28	6.10	6.10
FeO	0.80	0.73	0.88	0.60	0.60	0.01	0.04	0.10	0.01	0.01	0.04	0.04	0.08	0.01	0.11	0.11
MnO	0.03	0.03	0.03	0.03	0.03	0.64	0.11	0.27	0.12	0.27	0.17	0.25	1.39	0.07	2.71	2.71
MgO	0.47	0.40	0.41	0.33	0.64	0.11	0.27	0.12	0.27	0.17	0.25	1.39	0.07	2.71	2.71	2.71
CaO	1.71	1.86	1.66	2.00	2.03	1.73	1.64	1.55	0.33	0.69	0.77	1.01	1.41	2.28	5.23	5.23
Na ₂ O	3.17	3.30	3.29	3.38	3.21	3.64	3.70	3.82	5.12	7.27	3.70	3.09	3.30	4.46	3.25	3.25
K ₂ O	3.99	4.10	3.80	3.58	2.92	4.26	3.96	2.00	1.92	0.21	4.28	4.91	5.05	1.46	1.54	1.54
P ₂ O ₅	0.05	0.05	0.05	0.04	0.06	0.03	0.05	0.03	0.13	0.14	0.03	0.03	0.07	0.02	0.11	0.11
Rb	155.66	179.57	118.92	145.45	136.40	191.70	261.20	137.30	113.90	15.10	243.10	219.30	355.60	80.92	71.49	71.49
Ba	218.32	190.71	271.12	205.19	476.10	490.40	289.40	71.60	119.40	9.93	195.10	377.70	662.00	213.00	359.70	359.70
Sr	61.20	59.97	41.01	69.64	165.20	140.00	96.61	60.46	28.57	28.41	52.40	62.40	97.70	103.00	210.40	210.40
Zr	47.95	58.19	52.89	29.99	58.90	31.90	46.70	47.80	67.60	57.30	76.70	95.10	139.10	15.10	119.60	119.60
Y	20.83	30.17	28.70	29.05	15.30	30.35	33.32	71.06	25.70	21.88	41.30	44.70	24.82	20.94	20.94	20.94
La	5.33	7.46	3.94	6.29	15.79	10.33	10.04	3.83	10.26	4.32	24.12	33.61	23.87	9.89	22.62	22.62
Ce	11.29	15.88	8.48	13.23	31.97	21.27	20.81	8.07	24.03	10.81	52.42	74.51	50.48	21.67	45.10	45.10
Pr	1.21	1.72	0.93	1.41	3.86	2.44	2.43	0.94	2.93	1.34	6.46	8.57	6.08	2.53	5.16	5.16
Nd	4.86	6.91	3.77	5.71	14.01	9.35	9.22	3.86	10.60	4.86	24.07	31.87	23.45	9.76	19.35	19.35
Sm	1.34	1.85	1.08	1.72	2.88	2.49	2.47	1.71	3.04	1.81	5.99	7.22	5.38	3.01	3.92	3.92
Eu	0.28	0.34	0.24	0.47	0.84	0.77	0.59	0.19	0.22	0.09	0.55	0.60	0.63	0.35	0.92	0.92
Gd	1.64	2.28	1.59	2.31	2.77	2.87	3.05	3.09	3.20	2.11	6.23	7.17	5.88	3.19	3.86	3.86
Tb	0.37	0.52	0.41	0.54	0.50	0.59	0.66	0.95	0.67	0.53	1.19	1.25	1.09	0.64	0.59	0.59
Dy	2.71	3.78	3.36	3.88	3.25	4.01	4.43	8.10	4.20	3.43	7.47	7.36	7.04	3.97	3.44	3.44
Ho	0.67	0.94	0.89	0.92	0.77	0.92	0.97	2.05	0.83	0.68	1.66	1.56	1.62	0.81	0.70	0.70
Er	2.10	2.93	2.76	2.79	2.29	2.78	2.99	7.08	2.38	2.05	5.00	4.28	4.79	2.43	2.03	2.03
Yb	2.24	3.13	2.73	2.92	2.03	2.66	3.08	8.31	2.15	2.27	5.50	3.85	4.57	2.30	1.88	1.88
Lu	0.33	0.45	0.38	0.42	0.29	0.36	0.43	1.20	0.28	0.30	0.83	0.54	0.66	0.32	0.28	0.28
¹⁴³ Nd/ ¹⁴⁴ Nd	0.511932	0.511951	0.511927	0.511968	0.511947						0.512247	0.512143				
Error (2)	0.000060	0.000004	0.000004	0.000004	0.000004						0.000004	0.000002				
t (Ma)	10	10	10	10	10						10	10		10		
⁸⁷ Rb/ ⁸⁶ Sr	7.333554	8.637026	8.375155	6.031654	2.379854						13.43815	10.14899	10.49434			
⁸⁷ Sr/ ⁸⁶ Sr	0.76605	0.77149	0.78518	0.78308	0.763193						0.813269	0.782417	0.766507			
Error (2)	0.000034	0.000024	0.000044	0.000018	0.000010						0.000003	0.000003	0.000017			
E _{Nd(t)}	-13.7	-13.4	-13.8	-13.1	-13.2						-7.4	-9.4				
(⁸⁷ Sr/ ⁸⁶ Sr) _t	0.76501	0.77026	0.78399	0.78222	0.76285						0.81136	0.78098	0.76502			
T _{Zr} (C)	702	712	709	667	716	667	694	712	728	702	730	745	771	623	733	733
T _{Md} (C)	664	680	647	674	734	692	694	674	706	635	752	774	746	706	685	685

TABLE 1: continued: Major, trace and Sr, Nd isotopic data for North Himalayan Granites

DPP Granites																							
Data from Harris 2002						Data from JAK 2003						Data from Harris 2002						Data from JAK 2004					
Locality		Kouwo	T110	T111	Kouwo	T113	T114	Kouwo	DPP	JK3/25	DPP	T117	T118	Mabja	DPP + tur	T120	Mabja	T121	DPP	JK4/11b	JK4/12a	JK4/12b	Gomdre
Sample	Ma	DPP	14.4									9.8								14.5			
Rock type																							
SiO ₂		72.58	72.35	70.97	71.78	71.82	73.26	73.85	74.24	74.06	73.38	77.15	72.61	78.34									72.22
TiO ₂		0.29	0.29	0.39	0.35	0.27	0.11	0.10	0.08	0.12	0.10	0.10	0.29	0.02						0.10	0.29	0.02	0.27
Al ₂ O ₃		15.16	15.31	15.53	14.84	15.29	15.01	14.81	14.75	14.23	14.99	12.42	15.12	12.99						12.42	15.12	12.99	15.25
Fe ₂ O ₃		0.01	0.13	0.21	0.01	1.84	1.19	0.22	0.01	0.15	0.15	0.48	1.94	0.32						0.48	1.94	0.32	1.90
FeO		1.83	1.58	2.17	2.20			0.87	0.85	0.77	0.98												
MnO		0.03	0.03	0.03	0.03	0.03	0.02	0.03	0.03	0.02	0.03	0.01	0.03	0.05						0.01	0.03	0.05	0.03
MgO		0.94	0.83	1.10	0.99	0.71	0.21	0.35	0.31	0.26	0.37	0.32	0.74	0.05						0.32	0.74	0.05	0.70
CaO		1.55	1.37	1.62	1.84	1.62	0.73	0.94	1.02	1.05	0.95	0.20	1.59	0.35						0.20	1.59	0.35	1.56
Na ₂ O		3.16	3.34	2.97	3.46	3.48	3.31	3.34	3.54	3.65	3.43	3.56	3.53	3.91						3.56	3.53	3.91	3.12
K ₂ O		3.11	3.42	3.57	3.14	3.61	4.72	4.36	4.27	4.59	4.66	5.35	3.37	3.64						5.35	3.37	3.64	4.01
P ₂ O ₅		0.09	0.07	0.11	0.10	0.11	0.12	0.19	0.13	0.13	0.16	0.02	0.11	0.12						0.02	0.11	0.12	0.12
Rb		41.20	86.97	65.63	70.10	186.60	353.80	275.76	256.76	296.92	321.45	230.00	171.40	234.10						230.00	171.40	234.10	171.30
Ba		333.62	442.21	593.60	493.96	469.60	240.00	171.39	200.61	135.14	157.78	484.50	508.90	15.00						484.50	508.90	15.00	647.80
Zr		64.28	122.86	131.42	138.75	174.80	70.30	33.42	35.86	26.21	29.99	50.90	187.20	12.68						50.90	187.20	12.68	183.80
Sr		77.59	89.15	91.68	114.06	84.50	54.00	51.93	57.85	35.13	55.94	86.70	101.40	23.10						86.70	101.40	23.10	88.10
Y		9.53	12.62	16.40	16.69	11.72	12.00	7.68	8.34	5.51	9.28	18.86	10.43	4.19						18.86	10.43	4.19	14.80
La		13.17	16.25	25.88	22.46	23.42	17.43	9.16	8.85	6.26	11.13	30.22	25.23	1.51						30.22	25.23	1.51	28.69
Ce		28.24	34.72	55.01	47.70	47.98	35.59	19.25	18.56	12.98	23.32	64.34	51.56	3.33						64.34	51.56	3.33	58.89
Pr		3.09	3.78	5.94	5.14	5.51	4.16	2.04	1.96	1.35	2.46	7.68	5.97	0.40						7.68	5.97	0.40	6.95
Sm		12.20	14.88	23.32	20.15	20.67	15.00	7.84	7.40	5.08	9.22	28.59	22.32	1.40						28.59	22.32	1.40	25.69
Sr		2.73	3.35	5.05	4.40	4.45	3.63	1.96	1.76	1.25	2.33	6.33	4.79	0.54						6.33	4.79	0.54	5.33
Eu		0.56	0.74	1.00	0.93	1.03	0.53	0.24	0.17	0.19	0.23	0.89	1.12	0.04						0.89	1.12	0.04	1.16
Tb		2.38	2.96	4.33	3.49	3.82	3.49	1.76	1.64	1.15	2.11	4.75	4.07	0.51						4.75	4.07	0.51	4.32
Gd		0.35	0.45	0.62	0.57	0.51	0.60	0.29	0.27	0.20	0.35	0.66	0.53	0.12						0.66	0.53	0.12	0.58
Dy		1.76	2.29	3.11	2.93	2.37	3.10	1.47	1.46	1.01	1.77	3.44	2.31	0.67						3.44	2.31	0.67	2.57
Ho		0.33	0.44	0.59	0.47	0.40	0.57	0.25	0.27	0.18	0.31	0.67	0.36	0.12						0.67	0.36	0.12	0.43
Er		0.85	1.13	1.46	1.46	0.93	1.41	0.59	0.68	0.44	0.74	1.97	0.82	0.36						1.97	0.82	0.36	0.97
Yb		0.80	1.07	1.24	1.24	0.70	1.14	0.53	0.61	0.39	0.62	2.05	0.59	0.56						2.05	0.59	0.56	0.67
Lu		0.11	0.16	0.18	0.18	0.09	0.15	0.07	0.08	0.05	0.08	0.29	0.08	0.08						0.29	0.08	0.08	0.09
¹⁴³ Nd/ ¹⁴⁴ Nd		0.511697	0.511662	0.511623	0.511651	0.511623	0.511623	0.511623	0.511647	0.511697	0.511662	0.511662	0.511662	0.511932						0.511662	0.511662	0.511932	0.511932
Error (2)		0.000004	0.000004	0.000008	0.000006	0.000004	0.000004	0.000006	0.000006	0.000004	0.000004	0.000004	0.000004	0.000004						0.000004	0.000004	0.000004	0.000004
⁸⁷ Rb/ ⁸⁶ Sr		1.842822	2.035414	1.435788	1.452583	1.467104	1.467104	1.467104	20.81956	32.93795	31.17137	31.17137	31.17137	10						31.17137	31.17137	10	10
⁸⁷ Sr/ ⁸⁶ Sr		0.738483	0.738426	0.737695	0.737323	0.878522	0.878522	0.878522	0.853283	0.853197	0.854741	0.854741	0.854741	2.680178						0.854741	0.854741	2.680178	2.680178
Error (2)		0.000012	0.000014	0.000012	0.000012	0.000013	0.000013	0.000013	0.000030	0.000010	0.000012	0.000012	0.000012	0.740012						0.000012	0.000012	0.740012	0.740012
⁸⁷ Sr/ ⁸⁶ Sr _{init}		-13.7	-14.1	-13.5	-13.3	-19.5	-19.5	-19.5	-19.3	-18.3	-19.0	-19.0	-19.0	-13.5						-19.0	-19.0	-13.5	-13.5
⁸⁷ Sr/ ⁸⁶ Sr _t		0.73922	0.73814	0.73749	0.73712	0.87644	0.87644	0.87644	0.85033	0.84852	0.85031	0.85031	0.85031	0.73963						0.85031	0.85031	0.73963	0.73963
T _{2r} (C)		748	757	759	769	745	697	712	718	675	715	739	761	655						739	761	655	750
T _{1r} (C)		747	760	798	766	765	714	707	698	659	715	762	773	602						762	773	602	790

TABLE 2: U-Pb Age summary data for the Sakya dome granites

TABLE 2. U-Pb age summary data for the Sakya cone granites												
Granite	type	Sample	no. of analyses		Crystallisation age		no. of analyses		Inherited age		Analysis Instrument	
					Ma				Ma			
Kuday	AEG	JK3/05	8pt	cz	28.1	0.4	9pt	c	484	7	NuPlasma	
							2pt	c	868	40	NuPlasma	
Wing	AEG	JK4/07a	3pt	cz	27.5	1.0	7pt	c	492	5	NuPlasma	
Kua	AEG	JK4/08	16pt	ix	22.6	0.4	5pt	wz	563	12	Axiom	
Lijun	AEG	T03/33x	1pt	cz	25.4	6.0	6pt	c	499	16	NuPlasma	
Donggong	DPP	JK4/12a	10pt	wz	14.5	0.9	4pt	c	473	17	NuPlasma	
Gomdre	DPP	JK4/13g	12pt	ix	14.4	0.6					Axiom	
Mabja	DPP	T03/25i	4pt	oxm	8.8	0.2					Axiom	
			2pt	cx	11.2	0.5					Axiom	
c - concordant w - weighted mean i - lower intercept z - zircon m - monazite x - xenotime												

c - concordant

w - weighted mean

ix - lower intercept

z - zircon

m - monazite

x - xenotime

TABLE 3: Ar-Ar cooling ages for the North Himalayan Granites of the Sakya dome

Granite	type	Sample	lat	long	Mineral	Age (Ma)	2	previously published
Kuday	AEG	JK3/05	28 52.66	88 23.34	Biotite	10.5	0.8	
Kuday	AEG	JK3/06	28 53.28	88 23.40	Biotite	12.1	0.7	
Kuday	AEG	JK3/07	28 53.49	88 23.46	Biotite	11.5	0.4	
Kuday	AEG	JK3/08	28 53.75	88 23.41	Biotite	11.9	1.9	
Kuday	AEG	JK3/09	28 54.04	88 23.40	Biotite	12.5	0.4	
Kuday	AEG	JK3/10	28 54.28	88 23.41	Biotite	14.4	1.1	
Kuday	AEG	JK3/11	28 54.63	88 23.39	Biotite	13.5	0.3	
Kuday	AEG	JK3/12	28 55.10	88 23.37	Biotite	11.6	0.8	
Wing	AEG	JK4/07a	28 51.42	88 27.55	Biotite	14.1	0.7	
Kua	AEG	JK4/08	28 47.41	88 14.61	Biotite	13.5	0.2	
Kua	n/a	n/a	n/a	n/a	Muscovite	13.5	0.1	Lee et al. 2006
Kua	n/a	n/a	n/a	n/a	Biotite	13.6	0.1	Lee et al. 2006
Mabja	DPP	TO3/25i	28 39.28	87 52.83	Biotite	8.0	0.5	
Mabja	DPP	TO3/25ii	28 39.28	87 52.83	Biotite	8.6	0.3	
Mabja	DPP	n/a	n/a	n/a	Biotite	8.8	0.9	Maluski et al. 1988
Mabja	DPP	n/a	n/a	n/a	Muscovite	6.4	0.6	Maluski et al. 1988
Kouwo	DPP	JK3/25	28 45.72	88 04.72	Biotite	8.8	0.3	
Kouwo	DPP	n/a	n/a	n/a	Biotite	12.5	0.1	Lee et al. 2006
Kouwo	DPP	n/a	n/a	n/a	Muscovite	10.9	0.8	Zhang et al 2004
Kouwo	DPP	n/a	n/a	n/a	Biotite	11.4	1.1	Zhang et al 2004

n/a = not available

J = 0.00605 0.00003025 (1)

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