Oligocene–Miocene middle crustal flow in southern Tibet: geochronology of Mabja Dome

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Abstract: New U–Pb zircon, monazite, ⁴⁰Ar/³⁹Ar, and apatite fission track ages provide constraints on the timing of formation and exhumation of the Mabja Dome, southern Tibet, shed light on how this gneiss dome formed, and provide important clues on the tectonic evolution of middle crustal rocks in southern Tibet. Zircons from a deformed leucocratic dyke swarm yield a U–Pb age of 23.1 ± 0.8 Ma, providing the first age constraint on the timing of middle crustal ductile horizontal extension in the North Himalayan gneiss domes. Zircons and monazite from a post-tectonic two-mica granite yield ages of 14.2 ± 0.2 Ma and 14.5 ± 0.1 , respectively, indicating that vertical thinning and subhorizontal stretching had ceased by the middle Miocene. Mica ⁴⁰Ar/³⁹Ar ages from schists and orthogneisses increase structurally down-section from 12.85 ± 0.13 Ma to 17.0 ± 0.19 Ma and then decrease at the deepest structural levels to 13.29 ± 0.09 Ma. Micas from the leucocratic dyke swarm and post-tectonic two-mica granites yield similar ⁴⁰Ar/³⁹Ar cooling ages of 13.48 ± 0.13 to 12.84 ± 0.08 Ma. The low-temperature steps of potassium feldspar ⁴⁰Ar/³⁹Ar spectra yield ages of c. 11.0–12.5 Ma and apatite fission track analyses indicate the dome uniformly cooled below c. 115°C at 9.5 ± 0.6 Ma. Based on these data, calculated average cooling rates across the dome range from c. 40–60°C/million years in schist and orthogneiss and following emplacement of the leucocratic dyke swarm, to c. 350°C/million years following emplacement of the two-mica granites. The mylonitic foliation, peak metamorphic isograds, and mica ⁴⁰Ar/³⁹Ar chrontours are domed, whereas the low-temperature step potassium feldspar ⁴⁰Ar/³⁹Ar and apatite fission track chrontours are not, suggesting that doming occurred between 13.0 and 12.5 Ma and at temperatures between 370 and 200°C. Our new ages, along with field, structural and metamorphic data, indicate that the domal geometry observed at Mabja developed by middle-Miocene southward-directed thrust faulting upward and southward along a north-dipping ramp above cold Tethyan sediments. The structural, metamorphic and geochronologic histories documented at Mabja Dome are similar to those of Kangmar Dome, suggesting a common mode of occurrence of these events throughout southern Tibet. Vertical thinning and horizontal stretching, metamorphism, generation of migmatites, and emplacement of leucogranites in the domes of southern Tibet are synchronous with similar events in the Greater Himalayan Sequence that underlie the high Himalaya. These relations are consistent with previously proposed models for a ductile middle-crustal channel bounded above by the South Tibetan detachment system and below by the Main Central thrust in the High Himalaya that extended northward beneath southern Tibet.

The North Himalayan gneiss domes lie within the Tethys Himalaya south of the Indus–Tsangpo Suture Zone (ITSZ) and north of the South Tibetan detachment system (STDS) and crop out within the axis of the North Himalayan antiform (Fig. 1). These domes expose middle crustal meta-sedimentary rocks and orthogneisses that preserve contractional structures overprinted by moderate

temperature/pressure metamorphism, high strain structures developed during vertical thinning and horizontal stretching, partial melting, and emplacement of syn- and post-tectonic leucogranites (Burg *et al.* 1984; Chen *et al.* 1990; Lee *et al.* 2000, 2004; Aoya *et al.* 2005, 2006). While the structural, metamorphic and intrusive histories in these domes are well documented, their timing is not well known.

Fig. 1. Regional tectonic map of the central Himalaya orogen after Burchfiel *et al.* (1992) and Burg *et al.* (1984) showing location of the Mabja, Kangmar, Kampa and Malashan domes. GKT, Gyirong-Kangmar thrust fault system; ITSZ, Indus-Tsangpo suture zone; MBT, Main Boundary Thrust; MCT, Main Central Thrust; STDS, Southern Tibetan detachment system; YCS, Yadong cross-structure. Thrust faults represented by teeth on the hanging wall; normal faults by solid circle on the hanging wall. Inset (modified from Burchfiel *et al.* 1992; Tapponnier *et al.* 1982) shows location of regional tectonic map.

At Kangmar Dome (Fig. 1), mica $^{40}\text{Ar}/^{39}\text{Ar}$ ages showed that these rocks cooled below *c.* 370–335°C between *c.* 15 and 11 Ma, suggesting that D2 mylonitic vertical thinning and horizontal stretching deformation ended by middle Miocene (Lee *et al.* 2000). Apatite fission track ages indicated cooling below *c.* 115°C at *c.* 5.5 Ma, implying that these rocks were exhumed to the shallow crust by late Miocene (Lee *et al.* 2000). At Mabja Dome (Fig. 1), slightly deformed two-mica granites exposed west of the core of the dome (Maluski *et al.* 1988) yielded monazite U–Pb ages of 9.2 ± 0.9 Ma and 9.8 ± 0.7 Ma (Scharer *et al.* 1986) and biotite and muscovite collected from these granitic rocks and orthogneisses yielded disturbed $^{40}\text{Ar}/^{39}\text{Ar}$ spectra with total gas ages of 6–8 Ma (Maluski *et al.* 1988). The exact locations of these samples were not reported. More recently, U–Pb analyses of zircon, xenotime and monazite from granite intrusions to the north of Mabja and within Mabja yielded ages of 27.5 ± 0.5 Ma and 14.4 ± 0.1 Ma, respectively (Zhang *et al.* 2004).

To the south in the high Himalaya, the Greater Himalayan Sequence also exposes middle crust

including strongly deformed, moderate-temperature/pressure metasedimentary, orthogneissic and magmatic rocks, and both deformed and undeformed leucogranites (e.g. Le Fort *et al.* 1987; Hodges *et al.* 1988; Hubbard 1989; Burchfiel *et al.* 1992; Grujic *et al.* 1996, 2002; Murphy & Harrison 1999; Searle 1999a, b; Walker *et al.* 1999; Stephenson *et al.* 2001; Searle *et al.* 2003). These rocks preserve contractional structures overprinted by mylonitic fabrics and they are bounded by two major, north-dipping high-strain shear zones, the STDS normal fault at the top and the Main Central Thrust (MCT) fault at the base. In the high Himalaya, U–Pb, U–Th–Pb, and $^{40}\text{Ar}/^{39}\text{Ar}$ ages indicate three major events: (1) late Eocene to late Oligocene contraction-related burial and thermal re-equilibration (e.g. Vance & Harris 1999; Walker *et al.* 1999; Simpson *et al.* 2000); (2) early Oligocene to middle Miocene emplacement of multiple generations of both deformed and undeformed leucogranites (e.g. Noble & Searle 1995; Hodges *et al.* 1996, 1998; Edwards & Harrison 1997; Searle *et al.* 1997b; Wu *et al.* 1998; Harrison *et al.* 1999; Murphy & Harrison 1999; Searle 1999a, b; Walker *et al.* 1999;

Simpson *et al.* 2000; Daniel *et al.* 2003); and (3) an early to middle Miocene end to mylonitic deformation (e.g. Searle & Rex 1989; Hodges *et al.* 1992; Searle *et al.* 1992; Walker *et al.* 1999; Stephenson *et al.* 2001). In addition, the MCT and STDS shear zones were broadly active simultaneously during the early to middle Miocene (e.g. Hubbard & Harrison 1989; Hodges *et al.* 1992, 1996; Murphy & Harrison 1999; Walker *et al.* 1999; Simpson *et al.* 2000; Stephenson *et al.* 2001; Daniel *et al.* 2003; Searle *et al.* 2003).

Using structural, metamorphic, geochronologic and geophysical data, and thermal–mechanical models, workers have proposed that southward extrusion and erosion of ductile middle-crust bounded above by the STDS and below by the MCT can explain the exposure of the Greater Himalayan sequence in the high Himalaya (e.g. Grujic *et al.* 1996, 2002; Nelson *et al.* 1996; Searle 1999a, b; Beaumont *et al.* 2001; Hodges *et al.* 2001; Vannay & Grasemann 2001; Searle *et al.* 2003).

The parallel structural, metamorphic and intrusive histories of middle crustal rocks exposed within the Greater Himalayan Sequence, and within the North Himalayan gneiss domes, implies that these rocks represented a continuous section of middle crust beneath the high Himalaya and southern Tibet during the Oligocene and Miocene. To test this inference, geochronologic constraints on the timing of structural, metamorphic and intrusive events in the North Himalayan gneiss domes are needed. This paper presents new zircon and monazite U–Pb, mica $^{40}\text{Ar}/^{39}\text{Ar}$, and apatite fission track ages on structural, metamorphic and intrusive events at Mabja Dome, southern Tibet. Our new age data bracket the formation and exhumation of Mabja Dome, shed light on the mechanism by which this gneiss dome formed, and provide important clues on the tectonic evolution of middle crust rocks in southern Tibet.

Regional setting of the North Himalayan gneiss domes

The North Himalayan gneiss domes are exposed within the Tethyan Himalaya, approximately halfway between the ITSZ to the north and the STDS to the south (Fig. 1). The region is underlain by a Cambrian to Eocene miogeosynclinal sedimentary sequence deposited on the passive northern margin of the Indian continent (e.g. Gansser 1964; Le Fort 1975; Gaetani & Garzanti 1991). The Tethyan Himalaya is structurally complex, exhibiting Cretaceous to Quaternary reverse faults and folds (e.g. Le Fort 1975; Searle 1983; Burg & Chen 1984; Ratschbacher *et al.* 1994; Yin *et al.*

1994, 1999; Quidelleur *et al.* 1997; Searle *et al.* 1997a; Godin *et al.* 1999) and extensional structures (e.g. Molnar & Tapponnier 1975; Armijo *et al.* 1986; Mercier *et al.* 1987; Burchfiel *et al.* 1992; Ratschbacher *et al.* 1994) in a variety of orientations.

The North Himalayan gneiss domes consist of a core of metasedimentary rocks, gneisses and granitic rocks overlain by a mantle of sedimentary and low-grade metasedimentary rocks (e.g. Burg *et al.* 1984; Chen *et al.* 1990; Lee *et al.* 2000, 2004; Aoya *et al.* 2005, 2006). These rocks preserve evidence for a north–south contractional deformation event upon which a vertical thinning and horizontal stretching deformational event was superimposed during moderate temperature/pressure metamorphism, and intrusion of leucogranites. Along-strike, the exposure of the North Himalayan gneiss domes defines the North Himalayan antiform; the domes lie in the hanging wall of the north-dipping Gyirong–Kangmar thrust fault system (GKT) (Fig. 1).

Geology of Mabja Dome

Rock units

The Mabja Dome is a migmatitic orthogneiss mantled by Palaeozoic orthogneiss and metasedimentary rocks, that in turn are overlain by Triassic and Jurassic metasedimentary and sedimentary rocks (Figs 2 & 3). The grade of metamorphism ranges from sillimanite zone at the base to unmetamorphosed at the top (Lee *et al.* 2004). At the base of the section is a K-feldspar augen + biotite + plagioclase + quartz $\pm$ muscovite $\pm$ sillimanite $\pm$ garnet-bearing granitic migmatitic orthogneiss (og) that contains pockets and segregation banding of leucosomes and melanosomes suggesting partial melting. Structurally overlying unit og is a moderately well-exposed Palaeozoic orthogneiss and paragneiss complex (Pop) composed of dominantly K-feldspar granitic augen gneiss with numerous pendants of metasedimentary pelite (Figs 2 & 3). The metapelites include quartzite and coarse-grained, porphyroblastic schist, which range in metamorphic grade from garnet zone at the top, through kyanite- and staurolite zones in the middle, to sillimanite zone at the base. Structurally above Pop is a sequence of Palaeozoic schist, quartzite and marble comprising units Pls, Pm, Pus and Pq (Figs 2 & 3). Overlying unit Pq is an aerially extensive siliciclastic Triassic sedimentary sequence (Ts) which in turn is overlain by mudstones, sandstones and limestones of Jurassic age.

The orthogneiss and metasedimentary rocks were intruded by deformed amphibolite dykes, a variably

Fig. 2. Simplified geological map of Mabja Dome. Modified from Lee *et al.* (2004).

Fig. 3. Interpretative (a) NW-SE cross section and (b) SW-NE cross section of Mabja Dome. See Figure 2 for location. Modified from Lee *et al.* (2004).

deformed pegmatite and aplite dyke swarm, two undeformed biotite + muscovite granites, and an undeformed rhyolite porphyry dyke. Local, penetratively deformed, decimetre-wide amphibolite dykes, containing hornblende + plagioclase $\pm$ epidote $\pm$ garnet, are interlayered with schist and orthogneiss of unit Pop. A coarse- to medium-grained porphyritic to equigranular muscovite + K-feldspar + quartz $\pm$ biotite $\pm$ garnet-bearing pegmatite and fine-grained leucocratic aplite dyke swarm constitutes as much as 30–35% of the lower half of units Pop and og; this swarm first appears within the kyanite zone, and dramatically increases in volume downward toward the sillimanite zone (Figs 2 & 3). Two undeformed, medium- to coarse-grained, porphyritic two-mica granites, informally referred to as the Donggong and Kouwu granites (units Mdg and Mkg, respectively) were emplaced at deep structural levels into units Pls, Pop and og, and at shallower structural levels into unit Ts (Figs 2 & 3). Finally, an undeformed rhyolite porphyry dyke was emplaced across Ts and, in part, along the contact between the Kouwu granite and unit Ts.

Structural history

The Mabja Dome preserves evidence for three major deformational events, two older penetrative subhorizontal contractional and subhorizontal stretching events, and a younger doming event (Lee *et al.* 2004). D1, the oldest deformational event, is best exposed and dominant at the highest structural levels, exhibits bedding horizontally shortened into map- to mesoscopic scale, upright to inclined, open to tight, typically disharmonic F1 folds. At higher structural levels, the axial planar foliation to these folds, S1, is a spaced, pressure solution or crenulation cleavage that with increasing structural depth becomes a penetrative fine-grained slaty cleavage and finally a somewhat coarser-grained phyllitic cleavage.

Superimposed on D1 structural fabrics is D2 a high-strain deformational event that is manifested at higher structural levels as S2, a closely spaced to weakly penetrative crenulation cleavage developed at high angles to S1. S2 changes with increasing depth from a spaced axial planar cleavage, to open to tight folds of S1, to a penetrative axial planar cleavage, to isoclinal folds of S1. At structural levels below the garnet-in isograd, bedding and the S1 foliation have been transposed parallel to a mylonitic S2 foliation. Associated with the high-strain S2 foliation is a *c.* north–south stretching lineation, Ls2. The S2 mylonitic foliation is parallel to lithologic contacts and dips moderately NW on the NW flank of the dome and moderately

SW on the SW flank of the dome defining the domal geometry (Figs 2 & 3) (Lee *et al.* 2004).

Meso- and microscopic structures, such as strain shadows on porphyroblasts, tails on K-feldspar porphyroclasts, oblique quartz grain-shape foliations, shear bands, and small (centimetre-scale) normal faults, record the sense of shear associated with, and after development of, the high-strain S2 foliation within orthogneiss and metasedimentary rocks (Lee *et al.* 2004). Lee *et al.* (2004) suggested that the bulk shear strain history changes from dominantly coaxial during the high temperature, main phase of D2 deformation to dominantly top-S sense of shear during the low temperature, late phase of D2 deformation.

Subsequent to formation of D2 fabrics, the S2 foliation was domed into a doubly plunging, north–south elongate antiformal dome. The S2 mylonitic foliation dips moderately outward from the centre of the dome on the north, west and south flanks (Figs 2 & 3). Brittle structures are scarce and limited to two thrust faults of minor offset (tens of metres) and a 400–500 m dip-slip offset normal fault (Lee *et al.* 2004).

Metamorphic history

Microstructural textures indicate that peak metamorphism occurred after D1 deformation and prior to or during the D2 deformation. Peak metamorphism is defined by a prograde sequence of mineral assemblages that define a series of isograds (chloritoid-, garnet-, kyanite-, staurolite-, and sillimanite-in isograds) that increase towards the centre of the dome, are roughly concentric to the domal structure defined by the warped stratigraphy, and are parallel to the lithologic contacts and the S2 foliation. Based on mineral assemblages and quantitative thermobarometry, Lee *et al.* (2004) inferred temperatures and pressures of *c.* 475–530°C and *c.* 150–450 MPa for the chloritoid-zone and calculated temperatures that increase from $575 \pm 50^\circ\text{C}$ in the garnet zone to $705 \pm 65^\circ\text{C}$ in the sillimanite zone and pressures from garnet-, staurolite- and sillimanite-zone rocks that are constant at *c.* 800 MPa, regardless of structural depth.

Four important observations are apparent in the metamorphic petrology results (Lee *et al.* 2004). (1) The presence of pressures as high as 800 MPa (implying depths of *c.* 30 km) suggests that these rocks were thickened or buried. (2) Based on PT determinations outlined above, apparent isotherms can be drawn in which temperatures increase with structural depth, yielding a metamorphic field gradient of 10–60°C/km. (3) The apparent gradient in pressure between the chloritoid-in isograd and garnet-zone rocks is greater than expected,

indicating that these rocks were vertically thinned by *c.* 25–10% (horizontal stretching by a factor of *c.* four to ten). (4) Sillimanite-zone rocks exposed at the deepest structural levels yield calculated metamorphic pressures that are the same as, but higher temperatures than, garnet- and kyanite-zone rocks at shallower structural levels. Lee *et al.* (2004) interpreted this to indicate that peak pressures and temperatures occurred asynchronously, such that garnet-zone rocks reached peak metamorphic conditions at *c.* 30 km depth before kyanite-zone rocks reached peak conditions at the same depth, which in turn reached peak conditions before sillimanite-zone rocks, again at the same depth.

In summary, the Mabja Dome records two penetrative deformations, the second characterized by horizontal stretching and vertical thinning. At the deepest structural levels, emplacement of a pegmatite and aplite dyke swarm, development of migmatites, and initiation of doming was synchronous with ductile stretching. Lee *et al.* (2004) attributed development of migmatites to thermal re-equilibration and adiabatic decompression during regional extensional collapse, possibly enhanced by an unexposed granitic pluton beneath the core of the dome.

Geochronology and thermochronology

U–Pb geochronology

To determine the timing of the high-strain D2 deformational event, U–Pb geochronology by conventional thermal ionization mass spectrometry (TIMS) and sensitive high resolution ion microprobe (SHRIMP) techniques was completed on zircons and monazite from the post-tectonic Kouwu granite, and SHRIMP zircon analyses were performed on syn- to late-tectonic pegmatite dykes (Fig. 4, Table 1). Zircon and monazite were separated from 1–3 kg samples by standard gravity and magnetic techniques. Grains were hand-picked under alcohol for clarity, and lack of inclusions and cracks. TIMS analyses were performed on multigrain zircon and monazite fractions from the Kouwu granite. The fractions were spiked with a ^{205}Pb tracer solution and analysed at the TIMS facility at University of California, Santa Barbara, following procedures outlined in McClelland & Mattinson (1996). All of the TIMS zircon analyses are strongly discordant indicating the presence of significant inherited components in the multigrain fractions (Fig. 4). The SHRIMP technique was employed to improve spatial resolution and establish emplacement ages for the Kouwu granite and pegmatite dyke samples. Zircons selected for SHRIMP analysis were

mounted in epoxy and polished to expose grain centres. Cathodoluminescence (CL) images were used to characterize the grains and select spots for analysis (Fig. 5). Zircons were analysed for U–Pb on the SHRIMP–reverse geometry (RG) instrument at the Stanford University–United States Geological Survey Microanalytical Center (Palo Alto, California). A 30 μm diameter spot size was used for all analyses. The analytical routine followed Williams (1998) and data reduction utilized the SQUID program of Ludwig (2001).

Zircons were analysed from sample MD71, a two-mica pegmatite that is part of the pegmatite and aplite dyke swarm, is concordant to the S2 mylonitic foliation, does not possess a mesoscopic penetrative fabric, but does possess ribbon grains in thin section. These relations suggest that this dyke was emplaced syn- to late during D2 deformation. Zircons from this sample consist of oscillatory zoned cores surrounded by high U rims and tips (Fig. 5a). Cores yield older ages representative of inherited components. Excluding three older tip analyses interpreted to record mixing of inherited and new zircon and the youngest analysis interpreted to record younger disturbance, the tips yield a weighted $^{206}\text{Pb}/^{238}\text{U}$ mean age of 23.1 ± 0.8 Ma with a mean square of weighted deviate (MSWD) of 3.8, or 23.1 ± 1.6 Ma incorporating the uncertainty introduced by a MSWD = 3.8 (Fig. 4a). Penetrative D2 deformation was either continuing or was in its waning stages at these structural depths by this time.

Zircons from the undeformed Kouwu granite sample (MD86) contain inherited cores as well, as clearly indicated by the TIMS data (Figs 4b, c & 5b). Excluding four older rim and tip analyses interpreted to record mixing of inherited and new zircon and two younger analyses interpreted to record younger disturbance, tips from this sample yield a weighted $^{206}\text{Pb}-^{238}\text{U}$ mean age of 14.2 ± 0.2 Ma (MSWD = 0.5). A TIMS analysis of a single monazite yielded a U–Pb age of 14.5 ± 0.1 Ma. The TIMS monazite age is slightly older than the SHRIMP zircon age, perhaps indicating unresolved complexity in the monazite population. Our age for the Kouwu granite is not significantly different from the 14.4 ± 0.1 Ma age reported by Zhang *et al.* (2004). These data indicate that by 14.0–14.6 Ma, penetrative D2 deformation had ceased.

$^{40}\text{Ar}/^{39}\text{Ar}$ and apatite fission track thermochronology

$^{40}\text{Ar}/^{39}\text{Ar}$ and fission track were measured to characterize the exhumation history of the

Fig. 4. (a) Left, U–Pb Tera–Wasserburg concordia plot of all zircon data from sample MD71, a pegmatite dyke (SHRIMP analyses 9, 13 and 16 are not plotted). Right, plot of rim data used to calculate weighted mean ^{206}Pb – ^{238}U age. (b) Left, U–Pb concordia plot of all zircon data from sample MD86, the undeformed Kouwu two-mica granite. Location of TIMS analyses marked with squares (2σ ellipses are much smaller than squares). Right, plot of rim data used to calculate weighted mean ^{206}Pb – ^{238}U age (SHRIMP analysis 1 not plotted). TIMS monazite analysis plotted as solid ellipse (2σ) within square and denoted by ‘m’. (c) U–Pb Wetherill concordia plot of TIMS zircon and monazite data from sample MD86. See Figure 8 for location of samples. U–Pb plots were generated using Ludwig (1999).

high-grade rocks within the Mabja Dome. Two hornblende, 25 mica and eight potassium feldspar samples collected from deformed amphibolites, orthogneisses, schists and pegmatites, and

undeformed two-mica granites were analysed by $^{40}\text{Ar}/^{39}\text{Ar}$ methods. Seven apatite samples collected from orthogneisses, pegmatites and two-mica granites were analysed for fission track

Table 1. *U–Pb geochronologic data and apparent ages***A. SHRIMP U–Pb data**

Spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	f ²⁰⁶ Pb _c	²³⁸ U/ ²⁰⁶ Pb*	²⁰⁷ Pb/ ²⁰⁶ Pb*	²⁰⁶ Pb/ ²³⁸ U* (Ma)
Sample MD-71								
1 c	620	97	0.16	42	0.02	12.56 (1.5)	0.0558 (1.1)	495 (8)
2 c	1730	28	0.02	102	0.2	14.52 (1.6)	0.0573 (0.7)	428 (7)
3 c	2860	43	0.02	172	0.3	14.28 (1.5)	0.0580 (0.7)	435 (6)
4 t	5490	167	0.03	17	0.3	270.51 (1.6)	0.0488 (1.7)	#23.7 (0.4)
5 t	5990	144	0.02	19	0.4	267.84 (1.6)	0.0495 (4.5)	#23.9 (0.4)
6 t	5810	247	0.04	20	0.3	243.51 (2.6)	0.0488 (1.8)	26.3 (0.7)
7 t	3840	166	0.04	30	1.0	109.14 (1.5)	0.0552 (1.2)	58.2 (0.9)
8 t	5300	662	0.13	16	0.4	276.41 (1.7)	0.0499 (1.9)	#23.2 (0.4)
9 t	4160	273	0.07	14	8.2	263.31 (1.6)	0.1115 (3.8)	#22.4 (0.4)
10 t	6100	377	0.06	16	0.2	336.06 (1.5)	0.0482 (1.8)	19.1 (0.3)
11 t	8450	1470	0.18	89	1.1	81.10 (1.7)	0.0560 (0.7)	78.2 (1.3)
12 t	5090	527	0.11	15	0.05	291.59 (1.7)	0.0468 (1.9)	#22.1 (0.4)
13 c	610	482	0.82	131	0.8	3.98 (1.5)	0.0977 (0.5)	1434 (21)
14 c	250	184	0.77	16	0.2	13.14 (1.6)	0.0582 (1.7)	472 (7.4)
15 t	510	153	0.03	16	0.05	274.88 (2.1)	0.0469 (2.0)	#23.4 (0.5)
16 t	4460	113	0.03	16	6.1	238.56 (1.6)	0.0947 (5.3)	25.3 (0.4)
Sample MD86								
1 r	110	1	0.01	0.2	0.4	464.69 (4.3)	0.0495 (18.2)	#13.8 (0.6)
2 c	1440	220	0.16	101	0.004	12.29 (1.5)	0.0574 (0.9)	504 (8)
3 r	350	1	0.00	0.9	1.1	327.36 (2.4)	0.0555 (7.4)	19.4 (0.5)
4 r	1350	23	0.02	5	0.6	239.03 (1.7)	0.0509 (3.4)	26.8 (0.5)
5 t	5480	92	0.02	10	0.7	449.89 (1.6)	0.0516 (2.5)	#14.2 (0.2)
6 t	4190	257	0.06	8	0.5	478.99 (1.7)	0.0500 (3.0)	13.4 (0.2)
7 t	3780	64	0.02	8	3.9	426.67 (1.7)	0.0772 (3.8)	#14.5 (0.2)
8 t	470	19	0.04	0.8	0.5	498.85 (2.5)	0.0502 (8.3)	12.8 (0.3)
9 t	2740	86	0.03	6	0.7	414.03 (1.7)	0.0519 (3.5)	15.4 (0.3)
10 r	370	29	0.08	5	1.4	64.79 (2.4)	0.0594 (3.9)	97 (2.3)
11 t	1820	22	0.01	3	0.2	455.04 (1.8)	0.0476 (4.1)	#14.1 (0.3)
12 t	5500	101	0.02	10	0.1	456.52 (1.6)	0.0469 (2.5)	#14.1 (0.2)
13 t	1180	76	0.07	2	1.1	440.09 (1.9)	0.0549 (5.0)	#14.5 (0.3)
14 t	4550	86	0.02	9	0.1	452.85 (1.6)	0.0469 (2.7)	#14.2 (0.2)
15 r	3090	89	0.03	6	0.05	419.98 (1.7)	0.0467 (3.2)	15.3 (0.3)

Zircon SHRIMP analyses were performed on the SHRIMP-RG ion microprobe at the Stanford–United States Geological Survey Microanalytical Center at Stanford University. Spot abbreviations: number = grain number; c = core; r = rim; t = tip.

Pb*, radiogenic Pb; Pb_c, common Pb; f²⁰⁶Pb_c = 100(²⁰⁶Pb_c/²⁰⁶Pb_{total}).

*Calibration concentrations and isotopic compositions were based on replicate analyses of SL13 (238 ppm U) and R33 (419 Ma; Black *et al.* 2003). Reported SHRIMP ratios are not corrected for common Pb. Errors are reported in parentheses as percentages at the 1σ level. SHRIMP ages were calculated from ²⁰⁶Pb/²³⁸U ratios corrected for common Pb using the ²⁰⁷Pb method (see Williams 1998) and initial common Pb isotopic composition approximated from Stacey & Kramers (1975). Uncertainties in millions of years reported as 1σ. Ages annotated with a hash sign (#) were used in calculation of weighted mean ²⁰⁶Pb/²³⁸U ages.

B. TIMS U–Pb data: sample MD-86

Fraction size (μm)	Wt (mg)	U (ppm)	Pb* (ppm)	²⁰⁶ Pb/ ²⁰⁴ Pb*	²⁰⁶ Pb/ ²⁰⁷ Pb*	²⁰⁶ Pb/ ²⁰⁸ Pb*	²⁰⁶ Pb/ ²³⁸ U† (Ma)	²⁰⁷ Pb/ ²³⁵ U† (Ma)	²⁰⁷ Pb/ ²⁰⁶ Pb† (Ma)
a 50 × 50 × 80	0.50	2037	27	25,485	13.7491	12.3650	85.3 (0.2)	126.4 (0.3)	991 (1)
b 100 × 100 × 150	0.26	1791	30	27,877	13.9128	10.9274	107.6 (0.2)	155.6 (0.3)	968 (1)

(Continued)

Table 1. *Continued*

B. TIMS U–Pb data: sample MD-86

Fraction size (μm)	Wt (mg)	U (ppm)	Pb* (ppm)	$^{206}\text{Pb}/$ $^{204}\text{Pb}^*$	$^{206}\text{Pb}/$ $^{207}\text{Pb}^*$	$^{206}\text{Pb}/$ $^{208}\text{Pb}^*$	$^{206}\text{Pb}/$ $^{238}\text{U}^\dagger$ (Ma)	$^{207}\text{Pb}/$ $^{235}\text{U}^\dagger$ (Ma)	$^{207}\text{Pb}/$ $^{206}\text{Pb}^\dagger$ (Ma)
c 100 $\times$ 100 $\times$ 350	0.63	1651	43	12,671	16.4590	8.5824	164.3 (0.3)	195.4 (0.4)	590 (1)
d 175 $\times$ 175 $\times$ 400	0.60	1373	37	40,152	16.3649	14.6610	174.8 (0.3)	210.4 (0.4)	630 (1)
m 100 $\times$ 300 $\times$ 300	0.20	3723	30	563	13.8179	0.3280	14.6 (0.0)	14.5 (0.1)	4 (10)

TIMS analyses were performed at the University of California, Santa Barbara, following procedure described in McClelland & Mattinson (1996). Zircon fractions were abraded to 30 to 60% of original mass and washed in warm 3N HNO₃ and 3N HCl for 15 minutes each. All fractions were spiked with ^{205}Pb – ^{235}U tracer, and dissolved in a 50% HF $\gg$ 14N HNO₃ solution (zircon) or 12N HCl (monazite) within 0.5 ml SavillexTM capsules placed in 145 ml TFE TeflonTM lined Parr acid digestion bomb. Pb and U were combined and loaded with H₃PO₄ and silica gel onto single degassed Re filaments. Isotopic compositions of Pb and U were determined through static collection on a Finnigan-MAT 261 multicollector mass spectrometer utilizing an ion counter for collection of the ^{204}Pb beam.

Fraction: a, b, c, d designate conventional multigrain zircon fractions; m designates monazite fraction. Zircon fractions are non-magnetic on Frantz magnetic separator at 1.8 A, 15° forward slope, and side slope of 1°. Monazite fraction was magnetic on Frantz magnetic separator at 1.0 A, 20° forward slope, and side slope of 10°.

Pb*, radiogenic Pb.

*Reported ratios corrected for fractionation ($0.125 \pm 0.038\%$ /AMU) and spike Pb. Ratios used in age calculation were adjusted for 2 pg of blank Pb with isotopic composition of $^{206}\text{Pb}/^{204}\text{Pb} = 18.6$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.5$, and $^{208}\text{Pb}/^{204}\text{Pb} = 38.4$, 2 pg of blank U, $0.25 \pm 0.049\%$ /AMU fractionation for UO₂, and initial common Pb with isotopic composition approximated from Stacey & Kramers (1975) with an assigned uncertainty of 0.1 to initial $^{207}\text{Pb}/^{204}\text{Pb}$ ratio.

[†]Uncertainties reported as 2σ . Error assignment for individual analyses follows Mattinson (1987). An uncertainty of 0.2% is assigned to the $^{206}\text{Pb}/^{238}\text{U}$ ratio based on estimated reproducibility unless this value is exceeded by analytical uncertainties. Calculated uncertainty in the $^{207}\text{Pb}/^{206}\text{Pb}$ ratio incorporates uncertainty due to measured $^{204}\text{Pb}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios, initial $^{207}\text{Pb}/^{204}\text{Pb}$ ratio, and composition and amount of blank. Decay constants used: $^{238}\text{U} = 1.5513 \text{ E-10}$, $^{235}\text{U} = 9.8485 \text{ E-10}$, $^{238}\text{U}/^{235}\text{U} = 137.88$.

thermochronology. These samples were selected to provide age constraints on metamorphism and cooling histories across the dome along transects orientated approximately parallel and perpendicular to the Ls2 stretching lineation.

Fig. 5. Representative cathodoluminescence (CL) images of zircon analysed from samples (a) MD71 and (b) MD86. Ellipses indicate SHRIMP-RG analysis spots and the corresponding U–Pb ages ($\pm 1\sigma$ Ma).

For the $^{40}\text{Ar}/^{39}\text{Ar}$ samples discussed below, the estimated closure temperatures (assuming relatively rapid cooling rates) for hornblende, muscovite and biotite are $535 \pm 50^\circ\text{C}$ (Harrison 1981), $370 \pm 50^\circ\text{C}$ (Lister & Baldwin 1996), and $335 \pm 50^\circ\text{C}$ (Harrison *et al.* 1985; Grove & Harrison 1996), respectively. For the low-temperature steps of potassium feldspar spectra, we use an estimated closure temperature of $c. 200 \pm 50^\circ\text{C}$ (Harrison *et al.* 1995; Lee 1995). Weighted mean plateau ages (WMPA) are reported where more than 50% of the ^{39}Ar released in three or more contiguous steps is within 2σ error. For disturbed spectra, weighted mean ages (WMA) are reported where the spectrum is relatively flat, but does not meet the strict criteria for a plateau. The fission track age of a sample is usually interpreted as the time when the sample cooled below a closure temperature of $c. 120$ – 110°C (at a cooling rate of $c. 10^\circ\text{C}/\text{million years}$) (e.g. Naeser 1979) and is determined by measuring the density of fission tracks and the U concentration of the sample (Naeser 1976).

Analytical techniques and Ar/Ar data are available online at <http://www.geolsoc.org.uk/SUP18251>. A hard copy can be obtained from the Society Library. Argon isotopic ages, and fission track analyses are provided in Tables 2 and 3; age spectra are shown in Figures 6 and 7, sample localities are shown in Figure 8, and ages are projected on approximately north–south and east–west cross-sections in Figure 9.

Table 2. $^{40}\text{Ar}/^{39}\text{Ar}$ ages and calculated atmospheric $^{40}\text{Ar}/^{36}\text{Ar}$ ratios for muscovite and biotite samples

Sample	WMPA (Age $\pm$ 1 σ Ma)	WMA (Age $\pm$ 1 σ Ma)	TFA (Age $\pm$ 1 σ Ma)	Inverse Isochron (Age $\pm$ 1 σ Ma)	$^{40}\text{Ar}/^{36}\text{Ar}$ ($\pm$ 1 σ)	MSWD
Muscovite						
MD31B	—	12.79 $\pm$ 0.12	12.85 $\pm$ 0.12	12.85 $\pm$ 0.13	293.6 $\pm$ 4.3	1.59 < 2.63
MD48A	14.03 $\pm$ 0.14	—	13.96 $\pm$ 0.14	14.01 $\pm$ 0.14	298.2 $\pm$ 1.8	1.63 < 2.63
MD52	13.39 $\pm$ 0.12	—	11.56 $\pm$ 0.13	13.44 $\pm$ 0.12	289.3 $\pm$ 3.6	0.18 < 2.26
MD79	16.77 $\pm$ 0.15	—	16.33 $\pm$ 0.15	16.79 $\pm$ 0.16	290.1 $\pm$ 21.1	1.23 < 2.63
MD86	13.14 $\pm$ 0.05	—	13.09 $\pm$ 0.05	13.13 $\pm$ 0.06	296.0 $\pm$ 5.7	0.15 < 3.00
MD93	—	—	14.96 $\pm$ 0.14	14.23 $\pm$ 0.18	333.0 $\pm$ 9.0	1.93 < 3.00
MD97	13.54 $\pm$ 0.05	—	13.51 $\pm$ 0.06	13.54 $\pm$ 0.06	295.2 $\pm$ 1.9	0.71 < 2.41
MD100	—	16.27 $\pm$ 0.15	16.44 $\pm$ 0.15	16.12 $\pm$ 0.17	325.2 $\pm$ 19.6	2.12 < 3.00
MD47	—	16.99 $\pm$ 0.15	17.30 $\pm$ 0.16	17.09 $\pm$ 0.19	242.3 $\pm$ 61.7	0.02 < 3.83
MD49A	15.37 $\pm$ 0.14	—	15.52 $\pm$ 0.14	15.34 $\pm$ 0.14	307.1 $\pm$ 11.2	0.85 < 3.00
MD55	—	—	14.47 $\pm$ 0.13	14.78 $\pm$ 0.13	274.8 $\pm$ 3.1	0.58 < 3.00
MD71	13.48 $\pm$ 0.12	—	13.55 $\pm$ 0.13	13.48 $\pm$ 0.13	299.9 $\pm$ 7.1	0.97 < 2.63
MD64A	13.69 $\pm$ 0.12	—	13.66 $\pm$ 0.12	13.68 $\pm$ 0.13	295.6 $\pm$ 1.6	3.01 < 3.83
MD69B	13.33 $\pm$ 0.06	—	13.24 $\pm$ 0.06	13.29 $\pm$ 0.09	300.4 $\pm$ 7.4	1.52 < 3.00
Biotite						
MD33	—	15.51 $\pm$ 0.14	15.97 $\pm$ 0.14	15.52 $\pm$ 0.14	294.2 $\pm$ 1.5	1.08 < 2.15
MD39	—	17.60 $\pm$ 0.15	17.71 $\pm$ 0.16	17.66 $\pm$ 0.16	287.0 $\pm$ 3.0	0.75 < 2.63
MD48A	15.37 $\pm$ 0.14	—	15.30 $\pm$ 0.14	15.34 $\pm$ 0.22	312.2 $\pm$ 131.6	1.48 < 3.00
MD52	13.60 $\pm$ 0.12	—	13.57 $\pm$ 0.12	13.61 $\pm$ 0.12	292.7 $\pm$ 4.3	0.54 < 2.07
MD79	15.88 $\pm$ 0.15	—	14.92 $\pm$ 0.14	16.89 $\pm$ 0.75	231.2 $\pm$ 47.4	0.16 < 3.83
MD86	—	12.93 $\pm$ 0.06	13.00 $\pm$ 0.06	12.84 $\pm$ 0.08	340.0 $\pm$ 23.5	3.18 < 3.83
MD97	—	—	13.71 $\pm$ 0.06	13.48 $\pm$ 0.12	288.5 $\pm$ 29.3	83.44 > 3.83
MD37	13.49 $\pm$ 0.12	—	13.49 $\pm$ 0.15	13.58 $\pm$ 0.13	253.4 $\pm$ 27.9	0.35 < 3.00
MD71	—	13.60 $\pm$ 0.11	13.39 $\pm$ 0.11	13.59 $\pm$ 0.12	297.3 $\pm$ 11.4	3.34 < 3.83
MD64A	—	13.79 $\pm$ 0.13	13.72 $\pm$ 0.12	13.86 $\pm$ 0.13	247.9 $\pm$ 17.9	2.20 < 3.00
MD69A	—	13.48 $\pm$ 0.11	13.46 $\pm$ 0.11	13.49 $\pm$ 0.12	292.2 $\pm$ 14.9	1.87 < 3.83

WMPA, weighted mean plateau age; WMA, weighted mean age; TFA, total fusion age; MSWD, mean square of weighted deviates.

Hornblende Hornblende samples MD41 and MD49B from penetratively deformed amphibolites within the kyanite zone were analysed to provide an estimate for the age of peak metamorphism.

Both samples yield disturbed age spectra with double-gradient-type patterns suggesting incorporation of excess Ar, and are uninterpretable (Fig. 6).

Table 3. Apatite fission track analyses

Sample	No. grains	Standard track density ($\times 10^6 \text{ cm}^{-2}$)	Fossil track density ($\times 10^4 \text{ cm}^{-2}$)	Induced track density ($\times 10^4 \text{ cm}^{-2}$)	Chi squared prob. (%)	Central age (Ma)	Mean track length (μm)	Std. dev. (μm)
MD33	20	2.10 (3354)	2.03 (13)	82.5 (825)	94	8.3 $\pm$ 2.3		
MD39	20	2.10 (3354)	2.97 (19)	93.0 (595)	41	10.7 $\pm$ 2.6		
MD52	25	2.08 (3327)	9.38 (75)	326.9 (2615)	76	9.5 $\pm$ 1.2		
MD64	25	2.08 (3327)	4.75 (38)	187.8 (1502)	88	8.4 $\pm$ 1.4		
MD71	30	2.07 (3301)	6.10 (58)	175.8 (1672)	61	11.5 $\pm$ 1.6		
MD86	20	2.07 (3301)	16.6 (69)	598.3 (2489)	63	9.2 $\pm$ 1.2		
MD97	8	2.05 (3274)	14.9 (31)	50.3 (1046)	7	9.9 $\pm$ 2.3	14.4 $\pm$ 0.1 (65)	0.9

All apatite separates were prepared and analysed by A. Blythe at the University of Southern California. All ages are central ages (Galbraith & Laslett 1993). The conventional method (Green 1981) was used to determine errors on ages. Ages were calculated using $\text{zeta} = 320 \pm 9$ for dosimeter glass SRM 962a (e.g. Hurford & Green 1983). Numbers in parentheses represent total tracks counted or total lengths measured. The chi-squared test estimated the probability that individual grain ages for each sample belong to a single population with Poissonian distribution (Galbraith 1981).

Fig. 6. Hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra. Total fusion (TFA) and isochron (IA) ages given.

White mica Fourteen white mica samples from orthogneisses, schists, pegmatites and granites were analysed. Because most of the samples yield WMPA or WMA ages that are indistinguishable from their inverse isochron and/or total fusion ages and yield a trapped $^{40}\text{Ar}/^{36}\text{Ar}$ ratio that is not significantly different from the atmospheric ratio of 295.5 (Fig. 7, Table 2), we report their inverse isochron age (Figs 8 & 9). White mica ages from schists and orthogneisses increase down structural section from 12.85 ± 0.13 Ma at the top of garnet-zone rocks to 17.09 ± 0.19 Ma at the bottom of garnet-zone rocks, and then decrease at the deepest structural levels to 13.29 ± 0.09 Ma. This age pattern also holds true whether we report WMPA/WMA or total fusion ages. White micas from syn- to late-D2 pegmatites and post-tectonic two-mica granites yield ages of *c.* 13.4 Ma (range of 13.13 ± 0.06 Ma to 13.54 ± 0.06 Ma).

Biotite Like the muscovite samples, most biotites yield WMPA or WMA ages that are indistinguishable from their inverse isochron and total fusion ages, and yield a trapped $^{40}\text{Ar}/^{36}\text{Ar}$ ratio that is not significantly different from the atmospheric ratio of 295.5 (Fig. 7, Table 2); therefore, we report sample inverse isochron ages. Biotite ages also increase down structural section from 13.58 ± 0.13 Ma to 17.66 ± 0.16 Ma, and then decrease at the deepest structural levels to 13.49 ± 0.12 Ma. Biotites from the post-tectonic

two-mica granites yield ages of 12.84 ± 0.08 Ma to 13.61 ± 0.12 Ma.

Potassium feldspar Seven potassium feldspar samples from orthogneiss, pegmatite and two-mica granite were analysed (Fig. 7). Intermediate-depth orthogneiss samples MD33 and MD39 yield complex age spectra characterized by ages that climb steeply and erratically defining double-gradient patterns indicative of incorporation of excess argon. Ages older than 100 Ma occur at the high-temperature steps; these spectra are uninterpretable and are not shown in Figure 7. Migmatite sample MD64A yields an age spectrum with old apparent ages over the first *c.* 5% of ^{39}Ar released, suggesting incorporation of excess argon. Over the next *c.* 20% of the ^{39}Ar released, ages range from as young as *c.* 12.7 Ma to as old as *c.* 15 Ma, and then climb gradually to a maximum age of *c.* 17.7 Ma. Sample MD71, collected from a syn- to late-D2 pegmatite, yields a pattern indicative of excess argon in the first few low-temperature steps and then ages that climb slowly, but erratically, from *c.* 12.5 Ma to *c.* 14.4 Ma and then rapidly to 16.2 Ma over the last few high-temperature steps. Post-tectonic two-mica granite samples MD86 and MD97 yield simple spectra. At the lowest temperatures, these spectra exhibit little or no excess argon and low-temperature ages of *c.* 11.1–11.3 Ma that slowly climb to *c.* 13.0–13.1 Ma at high temperatures. Sample MD52 from a post-tectonic two-mica granite yields a double-gradient age spectrum pattern similar to orthogneiss samples, suggesting incorporation of excess argon.

Apatite Seven apatite separates from orthogneiss, granite and a pegmatite were analysed to constrain the low-temperature exhumation history (Figs 7 & 8). The fission track central ages (Galbraith & Laslett 1993) for the seven samples range from 8.3 ± 2.3 to 11.5 ± 1.6 Ma with 1σ uncertainty. The uncertainties are large on these ages because the apatites had relatively low concentrations of U. The seven fission track ages overlap at 1σ error, indicating that the dome uniformly cooled through *c.* 115°C at 9.5 ± 0.6 Ma, the mean age for all samples. Sample MD97 yields a mean track length of 14.4 ± 0.1 ($n = 65$) with a standard deviation of 0.9 indicating very rapid cooling.

Significance of geochronologic and thermochronologic results

Our U–Pb zircon and monazite measurements on the syn- to late-D2 leucocratic dyke swarm and post-D2 two-mica granites yield emplacement ages of 23.1 ± 0.8 Ma and *c.* 14.0–14.6 Ma, respectively. These data indicate that high-strain

Fig. 7. Muscovite (solid lines), biotite (dashed lines) and potassium feldspar (labelled Kspar) $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra. Shaded steps are those used in determining the weighted mean plateau age (WMPA) or weighted mean age (WMA). Apatite fission track ages are adjacent to solid diamonds. Inverse isochron (IA) and total fusion ages for each sample are listed in Table 2. Muscovite inverse isochron and apatite fission track ages are plotted on a simplified geological map in Figure 8; muscovite inverse isochron ages are projected onto NW–SE and NE–SW cross-sections in Figure 9.

Fig. 7. Continued.

Fig. 8. Simplified geological map of the Mabja Dome showing location of geochronology and thermochronology samples, U–Pb zircon, muscovite $^{40}\text{Ar}/^{39}\text{Ar}$, and apatite fission track ages. Chrontours for muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ inverse isochron ages from metamorphic and orthogneissic rocks shown as heavy lines; metamorphic isograds shown as thin lines. Solid and dashed lines indicate well-constrained and inferred locations, respectively, of chrontours and isograds. See Table 2 for muscovite inverse isochron ages and Figure 7 for age spectra.

D2 vertical thinning and horizontal stretching, synchronous with peak metamorphism and generation of migmatites, was continuing at 23.1 Ma and had ceased by c. 14.3 Ma.

Our white mica $^{40}\text{Ar}/^{39}\text{Ar}$ ages from metamorphic rocks range from 17.09 ± 0.19 Ma to

12.85 ± 0.13 Ma, and define approximately concentric chrontours centred on the core of the dome (Figs 8 & 9). The estimated $370 \pm 50^\circ\text{C}$ closure temperature for muscovite (Lister & Baldwin 1996) is somewhat higher than the minimum temperature at which quartz will deform ductilely.

Fig. 9. Muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ inverse isochron ages and chrontours (heavy lines), and metamorphic isograds (thin lines) projected onto the eastern parts of cross-sections A'–A'' and B'–B'' (see Figure 8). See Table 2 for muscovite inverse isochron ages and Figure 7 for age spectra.

Hence, D2 ductile deformation within Mabja Dome must have ended between 17 and 13 Ma, consistent with the *c.* 14.0–14.6 Ma emplacement age for the post-tectonic two-mica granite. White mica $^{40}\text{Ar}/^{39}\text{Ar}$ ages increase down structural section from 12.85 ± 0.13 Ma at the top of the garnet zone to 17.09 ± 0.19 Ma at the bottom of the garnet zone, and then decrease farther down structural section to *c.* 13.4 Ma in staurolite-zone rocks and deeper (Figs 8 & 9). At Kangmar Dome, Lee *et al.* (2000) documented an increase in mica $^{40}\text{Ar}/^{39}\text{Ar}$ age with structural depth, and attributed this to refrigeration from below due to underthrusting of a cold slab along the Gyirong–Kangmar thrust fault (GKT) (Fig. 1). Mabja Dome, like Kangmar Dome, lies in the hanging wall of the north-dipping GKT and, as for Kangmar, we suggest that the pattern of increasing mica $^{40}\text{Ar}/^{39}\text{Ar}$ ages with depth was caused by refrigeration of hot Mabja rocks by underthrust cold Tethyan sediments. In contrast to Kangmar, $^{40}\text{Ar}/^{39}\text{Ar}$ white mica ages in Mabja decrease at the deepest structural levels. This observation, together with U/Pb ages and field observations documenting that emplacement of the 14.0–14.6 Ma two-mica granites post-date D2 deformation, indicates that middle Miocene refrigeration at the deepest structural levels was likely overprinted by a reheating event at *c.* 14.3 Ma followed by rapid conductive cooling. Finally, low temperature potassium feldspar $^{40}\text{Ar}/^{39}\text{Ar}$ and apatite fission track data yield

uniform ages demonstrating that the dome symmetrically cooled between $200 \pm 50^\circ\text{C}$ and $115 \pm 5^\circ\text{C}$ from *c.* 12.5 Ma to 9.5 ± 0.6 Ma (Figs 7 & 8).

Calculated cooling rates across the dome based on our zircon and monazite U–Pb, mica and potassium feldspar $^{40}\text{Ar}/^{39}\text{Ar}$, and apatite fission track data are shown in Figure 10. The syn- to late-tectonic pegmatite exhibits rapid average cooling rates of 40–60°C/million years following emplacement at 23.1 ± 0.8 Ma and a zircon closure temperature of *c.* 750°C (Cherniak & Watson 2003), to a temperature of $115 \pm 5^\circ\text{C}$ at 9.5 ± 0.6 Ma (Fig. 10). Initial rapid cooling of the pegmatite at a rate of *c.* 350°C/million years followed by slower cooling at 6°C/million years might be a more reasonable initial cooling history (Fig. 10). Shallow (garnet zone), intermediate (garnet and kyanite zone) and deep (sillimanite zone) structural levels also yield rapid cooling rates of 45–60°C/million years from *c.* 370–335°C to 120–110°C between 17.09 ± 0.19 Ma and 9.5 ± 0.6 Ma (Fig. 10). Finally, the two-mica granites cooled rapidly at rates of *c.* 350°C/million years from a zircon closure temperature of *c.* 750°C at *c.* 14.3 Ma to *c.* 370°C at *c.* 13.4 Ma (Fig. 10), suggesting conductive cooling and possibly exhumation. Continued cooling of the two-mica granites from $370 \pm 50^\circ\text{C}$ at *c.* 13.4 Ma to $115 \pm 5^\circ\text{C}$ at 9.5 ± 0.6 Ma was slower, but still high, at a rate of 60°C/million years; this is essentially the same cooling rate and age range as documented in the metasedimentary and orthogneissic rocks.

Fig. 10. Temperature-time plot showing cooling histories for geochronology and thermochronology samples from Mabja Dome.

Deep, intermediate and shallow structural levels, the pegmatite dyke swarm, and the post-tectonic granite intrusions yield nearly identical rapid cooling rates of *c.* 45–60°C/million years from $370 \pm 50^\circ\text{C}$ to $115 \pm 5^\circ\text{C}$ between 17.09 ± 0.19 to 13.29 ± 0.09 Ma and 9.5 ± 0.6 Ma, which are most likely the result of a combination of refrigeration and exhumation during thrust faulting and erosion. If we assume a linear, steady-state geothermal gradient of 30°C/km for the crust, these cooling rates imply an exhumation rate of *c.* 1.5–2.0 mm/a. This assumption is probably reasonable in areas of slow denudation, but may not be correct where the thermal structure of the crust has been perturbed by advection (Mancktelow & Grasemann 1997; Moore & England 2001), implying that the exhumation rate we calculate may be a minimum.

Discussion

Formation of Mabja Gneiss Dome

The mechanism by which Himalayan gneiss domes form is typically attributed to one or more of three

processes: metamorphic core complex-type extension, diapirism, or duplex formation (Burg *et al.* 1984; Le Fort, 1986; Le Fort *et al.* 1987; Chen *et al.* 1990; Lee *et al.* 2000, 2004). Based on field, structural and metamorphic petrology data from Mabja, Lee *et al.* (2004) advocated a doming mechanism driven, at least in part, by buoyant migmatite diapirs generated during adiabatic decompression synchronous with D2 ductile extensional collapse and possibly enhanced by a proposed buoyant granitic body at depth. In contrast, Lee *et al.* (2000), building on the work of Burg *et al.* (1984) and Chen *et al.* (1990), concluded that the domal geometry in Kangmar was the result of hot middle crust thrust up and over a north-dipping ramp along the GKT; thermochronologic data were the key dataset to this interpretation. Without similar thermochronology, Lee *et al.* (2004) could not exclude the possibility that a similar regional contractional and/or extensional event contributed to the growth of the domal form in Mabja.

Our new ages, in conjunction with field, structural and metamorphic data from Lee *et al.* (2004) and regional field relations, provide new insight into the formation of Mabja Dome (Fig. 11).

The S2 foliation, M1 isograds, and muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ chrontours are domed (Figs 8 & 9), whereas the low-temperature-step potassium feldspar $^{40}\text{Ar}/^{39}\text{Ar}$ and apatite fission track chrontours are not. This observation means that an episode of

doming occurred at temperatures below *c.* 370°C (the estimated blocking temperature for muscovite) after *c.* 15 Ma, the youngest muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ chrontour that appears to be domed, but above *c.* 200°C (the estimated blocking temperature for

Fig. 11. Interpretative, schematic, evolutionary north-south cross-sections across the Himalaya and southern Tibet at the approximate longitude of Mabja Dome. Sections show major structures, middle-crustal migmatites, D2 horizontal stretching fabrics, cold and strong Indian crust, the GKT, and post-tectonic plutons. See text for details. Modified from Hauck *et al.* (1998), Wu *et al.* (1998), Makovsky *et al.* (1999), Lee *et al.* (2000) and Beaumont *et al.* (2004). GKT, Gyirong-Kangmar thrust fault system; MCT, Main Central Thrust; MD, Mabja Dome; MHT, Main Himalayan Thrust; STDS, Southern Tibetan detachment system.

low-temperature-step potassium feldspar) and before *c.* 12.5 Ma, the age of uniform cooling recorded in the low-temperature steps in potassium feldspar spectra (Fig. 7).

The structural, metamorphic and cooling histories recorded within the Mabja rocks are similar to those at Kangmar, but the intrusive histories are different (cf. Burg *et al.* 1984; Chen *et al.* 1990; Lee *et al.* 2000, 2004), and thus we suggest that Mabja formed by a mechanism similar to that proposed for Kangmar (Lee *et al.* 2000), but with some modifications (Fig. 11). As at Kangmar, normal faults were not documented that could have accommodated the middle crustal penetrative D2 vertical thinning and horizontal stretching deformation in the Mabja region. To maintain strain compatibility, we follow Lee *et al.*'s (2000) interpretation that these D2 deformational fabrics exposed in the core of the dome were accommodated at shallow crustal levels to the south by normal-sense (top-to-north) slip along the STDS (Fig. 11a). This interpretation implies that normal slip along the STDS must have been ongoing by at least 23 Ma, consistent with the inference that ductile shear along the STDS pre-dates *c.* 17 Ma (e.g. Murphy & Harrison 1999; Searle *et al.* 2003) (Fig. 11a). Moreover, our interpretation suggests that the D2 vertical thinning and horizontal stretching fabrics observed within the North Himalayan gneiss domes were accommodated by southward extrusion or channel flow of the middle crust beneath southern Tibet (e.g. Grujic *et al.* 1996, 2002; Searle 1999a,b; Beaumont *et al.* 2001, 2004) by at least early Miocene. The most important criterion for channel flow is low viscosity (Beaumont *et al.* 2001, 2004), which Beaumont *et al.* (2004) postulated can be achieved by a small amount of partial melt. The generation of migmatite and emplacement of a leucocratic dyke swarm, both syntectonic with the development of D2 ductile extensional deformation fabrics (Lee *et al.* 2004), could have provided the partial melt necessary to lower the viscosity and initiate channel flow in the middle crust of southern Tibet.

Lee *et al.* (2004) argued that the close spatial and temporal relations among metamorphism, partial melting, emplacement of the dyke swarm, and D2 vertical thinning and horizontal stretching in Mabja indicated that doming was, in part, driven by buoyant migmatite diapirs that were generated by adiabatic decompression during extensional collapse. We can test this high-temperature (*c.* 500–700°C) doming mechanism by comparing the degree of fold tightness exhibited by the domed high-temperature S2 foliation and metamorphic isograds with the domed lower temperature (*c.* 370°C) muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ chrontours. If the migmatite diapir mechanism is correct, then the

domal geometry defined by the S2 foliation and metamorphic isograds should be tighter than that defined by the $^{40}\text{Ar}/^{39}\text{Ar}$ chrontours. Both map and cross-sectional views of the S2 foliation, metamorphic isograds and $^{40}\text{Ar}/^{39}\text{Ar}$ chrontours show they are subparallel and do not exhibit a difference in degree of folding (Figs 8 & 9), indicating that the entire doming history in Mabja must have occurred after the rocks had cooled below *c.* 370°C.

Between *c.* 17 Ma and *c.* 15 Ma, hot Mabja rocks were captured in the hanging wall of the north-dipping GKT and thrust southward up and over cold Tethyan sediments, resulting in refrigeration from below, an increase in muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ ages with structural depth, rapid cooling (40–60°C/million years) due to refrigeration and exhumation, and 'freezing-in' middle-crustal channel flow fabrics (Fig. 11b) (cf. Lee *et al.* 2000). Continued movement of these rocks up and over a north-dipping ramp along the thrust fault between *c.* 13 Ma and *c.* 12.5 Ma resulted in passive doming of M1 isograds, S2 foliations, and mica chrontours, and continued rapid cooling at a rate of 45–60°C/million years (Fig. 11c). In contrast to Kangmar, muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ ages observed at the deepest structural levels in Mabja decrease to *c.* 13–14 Ma, suggesting these rocks were reheated above the closure temperature for muscovite at *c.* 15–16 Ma and then rapidly cooled as a consequence of conduction and exhumation. We propose that a granite, below the present level of exposure and similar in composition and age to the two post-tectonic granites exposed, was the source of this additional heat (Fig. 11c). Symmetric cooling of the dome from *c.* 200°C to 115°C between *c.* 12.5 and 9.5 Ma implies either cessation of thrust faulting and rapid exhumation (45–60°C/million years) due to erosion following thrust faulting, or continued slip along a subhorizontal portion of the GKT and erosion.

Changes in horizontal stress, shear traction at the base, rheology and/or surface height (e.g. Dahlen 1984, 1990; Davis *et al.* 1983; Platt 1986; England & Molnar 1993) can explain the transition from extension to compression in an orogenic belt. We (Lee *et al.* 2000, 2004) speculated that increased friction along the Main Himalayan thrust (MHT) beneath the North Himalayan gneiss domes (Fig. 11) led to the change from middle crustal subhorizontal stretching deformation to contraction and formation of the GKT. Alternatively, Beaumont *et al.*'s (2001, 2004) thermal-mechanical model of strong crust injected into and underthrusting a middle crustal channel results in the development of a north-dipping frontal ramp. The north-dipping ramp along the MHT proposed by Hauck *et al.* (1998) may be the leading edge of Beaumont *et al.*'s (2001, 2004)

north-dipping frontal ramp (Fig. 11b). The cold crustal ramp provides an explanation for the development of a north-dipping ramp along the GKT proposed by Lee *et al.* (2000) (Fig. 11b). This hypothesis is more appealing because it does not require changes in stress, shear traction or rheology in a contractional deformation zone that is well established.

A baffling feature of the North Himalayan gneiss domes is their map geometries, which vary along-strike from elongate north–south (Kangmar), to elongate east–west (Kampa), to asymmetric with both north–south and east–west elongate components (Mabja) (Fig. 1). To explain these variations, we propose that the shape and size of the leading edge of the cold, strong frontal ramp varies along its length. We envision that a north–south elongate, narrow footwall underthrust middle crust rocks now exposed at Kangmar, whereas an asymmetric, north–south and east–west elongate footwall underthrust Mabja. If this hypothesis is correct, then in map view the leading edge of the frontal ramp would be characterized by a sinuous pattern in which the enveloping surface strikes east–west, and lateral ramps would be developed under the east and west flanks of the domes.

There are striking parallel events among structural, metamorphic and cooling histories preserved at Mabja (Lee *et al.* 2004, this work) and Kangmar (Burg *et al.* 1984; Chen *et al.* 1990; Lee *et al.* 2000): (1) D1 deformation characterized by north–south contraction resulting in folding and thickening; (2) thermal re-equilibration of middle crustal rocks leading to peak metamorphism; (3) high-strain D2 vertical thinning and north–south horizontal stretching broadly synchronous with peak metamorphism; and (4) development of a domal geometry defined by lithologic contacts, the S2 mylonitic foliation, metamorphic isograds, and $^{40}\text{Ar}/^{39}\text{Ar}$ muscovite chronotours. Moreover, both domes lie in the hanging wall of the GKT (Fig. 1) and their domal geometries are ascribed to movement of the hangingwall of this thrust fault. Conversely, the migmatites, leucocratic dyke swarm, and post-tectonic granites of the Mabja Dome (Lee *et al.* 2004) were not found at Kangmar Dome (Lee *et al.* 2000), perhaps because Mabja exposes deeper structural levels than Kangmar. The similarities between these two domes suggest that the physical processes that produced them may be the same for other North Himalayan gneiss domes and are, therefore, of regional extent.

Miocene middle crustal flow in southern Tibet

To the south of the North Himalayan gneiss domes, the Greater Himalayan Sequence in the high

Himalaya is underlain by middle crust composed of strongly deformed high-grade metasedimentary (peak conditions of 6–7 kbar and *c.* 600°C), orthogneissic and migmatitic rocks, and both deformed and undeformed leucogranites (e.g. Le Fort *et al.* 1987; Hodges *et al.* 1988; Hubbard 1989; Burchfiel *et al.* 1992; Grujic *et al.* 1996, 2002; Murphy & Harrison 1999; Searle 1999*a,b*; Walker *et al.* 1999; Stephenson *et al.* 2000; Searle *et al.* 2003). Numerous U–Pb and U–Th–Pb geochronologic ages on zircon, monazite and other uranium-bearing accessory phases, together with $^{40}\text{Ar}/^{39}\text{Ar}$ ages on micas, constrain the timing of structural, metamorphic and intrusive events recorded in these rocks. Contraction-related burial and peak Barrovian-type metamorphism (thermal re-equilibration) have been bracketed between 37 Ma and 28 Ma (Vance & Harris 1999; Walker *et al.* 1999; Simpson *et al.* 2000); these data also provide a minimum age for north–south contraction. U–Pb ages, in conjunction with petrographic and trace-element partitioning observations, from the Namche migmatites in the Everest region are interpreted as indicating anatexis at 25.4–24.8 Ma (Viskupic & Hodges 2001). The multiple generations of both deformed and undeformed leucogranites in the high Himalaya vary in age from 31.6 Ma to 12.5 Ma (Noble & Searle 1995; Hodges *et al.* 1996, 1998; Edwards & Harrison 1997; Searle *et al.* 1997*b*; Wu *et al.* 1998; Harrison *et al.* 1999; Murphy & Harrison 1999; Searle 1999*a,b*; Walker *et al.* 1999; Simpson *et al.* 2000), and their production is attributed to crustal melting either as a consequence of decompression during exhumation (e.g. Harris *et al.* 1993; Harris & Massey 1994) or as a consequence of heat generated during shear stress along the Himalayan decollement (e.g. Le Fort *et al.* 1987; Harrison *et al.* 1997). Muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ ages indicate that most of the high Himalaya metasediments cooled below *c.* 370°C by between 22–20 Ma and 17–15 Ma signifying that mylonitization ceased at this time or soon thereafter (Searle & Rex 1989; Hodges *et al.* 1992; Searle *et al.* 1992; Walker *et al.* 1999; Stephenson *et al.* 2001). Finally, the combination of field, structural and geochronologic observations indicate that movement along the MCT and STDS shear zones, that bound the Greater Himalayan Sequence below and above, respectively, was broadly simultaneous at *c.* 22–13 Ma (Hubbard & Harrison 1989; Hodges *et al.* 1992, 1996; Murphy & Harrison, 1999; Walker *et al.* 1999; Simpson *et al.* 2000; Stephenson *et al.* 2001; Daniel *et al.* 2003; Searle *et al.* 2003).

The structural, metamorphic, anatectic, intrusive and geochronologic histories in Mabja (Lee *et al.* 2004, this study), Kangmar (Burg *et al.* 1984; Chen *et al.* 1990; Lee *et al.* 2000), and Malashan

(Burg *et al.* 1984; Aoya *et al.* 2006) are similar to those recorded in the Greater Himalayan Sequence, suggesting that during the late Oligocene to early Miocene high-grade middle-crustal metasedimentary and orthogneissic rocks, cross-cut by anatectic melts and leucogranites, were once continuous from beneath the high Himalaya northward beneath southern Tibet (Fig. 11a). Exposures of the Greater Himalayan Sequence have been interpreted as the leading edge of an eroding and southward-extruding tabular or wedge-shaped body of ductile middle-crustal rocks bounded above by the STDS and below by the MCT (e.g. Grujic *et al.* 1996, 2002; Nelson *et al.* 1996; Searle 1999a, b; Beaumont *et al.* 2001, 2004; Hodges *et al.* 2001; Vannay & Grasemann 2001; Searle *et al.* 2003). Beaumont *et al.*'s (2001, 2004) thermal–mechanical models predict that channel flow within the middle crust will develop in the Himalayan orogen if low-viscosity partial melt is present in the middle crust, differential lithostatic pressure is established across the orogen, and surface denudation occurs along the southern flank of the Himalaya. We suggest that the hot and weak middle crustal rocks now exposed in the core of the North Himalayan gneiss domes represent the interior of such a middle-crustal channel. Present-day exposures of mid-crustal rocks in the gneiss dome core signify that a piece of the middle-crustal channel was excised in southern Tibet. Our interpretation that exhumation of these high-grade rocks was related to their movement up and over a strong, crustal frontal ramp (Beaumont *et al.* 2001, 2004) and into the hanging wall of the GKT during the middle Miocene provides a mechanism by which this piece of the flowing middle crustal channel was cut out (Fig. 11b). New and previously published geochronologic data indicate that ductile flow was synchronous in the core of Mabja Dome and in the high Himalaya, implying that the middle crust throughout southern Tibet and the high Himalaya was, in general, flowing southward by early Miocene times.

Studies along the southern flank of the Himalaya have concluded that the deformation field of the channel flow tunnel or extruding wedge can be described as shear along the MCT and STDS (e.g. Hodges *et al.* 1993), simple shear distributed throughout the wedge (Grujic *et al.* 1996), or general shear flow concentrated along the boundaries of the wedge with pure shear in the centre (Grasemann *et al.* 1999). Although sparse, kinematic data from the North Himalayan gneiss domes (e.g. Chen *et al.* 1990; Lee *et al.* 2000, 2004; Aoya *et al.* 2005, 2006) imply that these middle-crustal rocks were probably deformed along the upper part of the deforming wedge or channel. Detailed quantitative kinematic studies may resolve where in the channel these rocks were deformed.

Geophysical observations, including short-wave-length gravity anomalies (Jin *et al.* 1994), and the coincidence of high electrical conductivity, middle-crustal low velocities, and reflection bright spots (Chen *et al.* 1996; Makovsky *et al.* 1996; Nelson *et al.* 1996; Alsdorf & Nelson 1999), suggest that the middle crust beneath southern Tibet is currently hot, partially molten, and weak. The migmatites and leucogranites in the high Himalaya and in the North Himalayan gneiss domes are exposures of the once hot and weak Oligocene–Miocene middle crust (Nelson *et al.* 1996; Searle *et al.* 2003; Lee *et al.* 2004).

Conclusions

New isotopic ages from the Mabja Dome reveal a late Oligocene to early Miocene history of ductile vertical thinning and horizontal stretching, peak metamorphism, migmatization and emplacement of a leucocratic dyke swarm, early to middle Miocene south-vergent thrust faulting resulting in doming, and post-tectonic emplacement of middle Miocene two-mica granites. Our 23.1 ± 0.8 Ma U–Pb zircon age from a deformed leucocratic dyke are the first to constrain the timing of ductile extension, metamorphism and migmatization within the North Himalayan gneiss domes. Mica $^{40}\text{Ar}/^{39}\text{Ar}$ and apatite fission track cooling ages indicate rapid cooling and doming during the middle Miocene. Rapid cooling is attributed to both refrigeration from below and exhumation. The domal geometry observed at Mabja is solely ascribed to tectonically driven south-vergent thrust faulting. The similar structural, metamorphic, intrusive and timing histories at Mabja Dome and in the Greater Himalayan sequence imply that during late Oligocene to early Miocene times, high-grade metasedimentary rocks and orthogneissic rocks, intruded by migmatites and leucogranites, were continuous in the middle crust from beneath the High Himalaya northward to beneath southern Tibet. These middle crustal rocks have been interpreted as a southward-flowing middle crustal channel, with the Greater Himalayan Sequence defining the eroding and extruding leading edge. In southern Tibet, a slice of the southward-flowing middle crustal channel was excised by south-vergent thrust faulting during the middle Miocene, explaining the present-day exposures of high-grade rocks in the cores of the North Himalayan gneiss domes. Excisement via thrust faulting was broadly simultaneous with normal slip along the STDS and reverse slip along the MCT.

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