Research article

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Nature, age and emplacement of the Spongtang ophiolite, Ladakh, NW India

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Abstract: The Spongtang ophiolite (Ladakh, NW India) constrains the nature of oceanic lithosphere before Indo-Asia collision and key stages in the development of the Himalayas. We report whole-rock 40 Ar/ 39 Ar and *in situ* zircon 238 U– 206 Pb ages from its crustal and upper and lower mantle sequences. Major and trace elements from harzburgite minerals suggest that the ophiolite formed at a mid-ocean ridge-type spreading centre, whereas published spinel compositions from Spongtang dunites are consistent with a suprasubduction-zone setting. Rare earth element-in-two-pyroxene thermometry for the harzburgite yields $1058 \pm 13^{\circ}$ C whereas temperature from solvus-based two-pyroxene and olivine–spinel thermometry is lower (to 656° C). The distribution suggests that the mantle section of the ophiolite cooled at rates of 100° Ma⁻¹ or slower. Based on ages, major and trace element geochemistry, and geospeedometric estimates, we model the origin of the Spongtang ophiolite as forming within a mid-ocean ridge-type spreading centre with a spreading rate >2 cm a⁻¹ in the Neotethyan Ocean, possibly from the Late Triassic to Jurassic. By the Early Cretaceous, the ridge experienced increasing influence of subduction beneath the Spongtang oceanic lithosphere owing to a subduction polarity reversal. Based on 238 U– 206 Pb ages of the youngest Cenozoic zircon grain, latest obduction occurred between 64.3 ± 0.8 and 42.4 ± 0.5 Ma, in accordance with 56.7 ± 5.2 Ma whole-rock 40 Ar/ 39 Ar ages.

Supplementary material: Excel files with details of electron microprobe and inductively coupled plasma mass spectrometry (ICP-MS) analyses, argon isotopic whole-rock and secondary ion mass spectrometry (SIMS) analyses, and the T_{REE} calculations, including an inversion diagram showing regression through measured REE distributions in cpx and opx (from Liang *et al.* 2013), are available at https://doi.org/10.6084/m9.figshare.c.4261856

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During the mid-Mesozoic to Eocene, the Indian subcontinent moved over 60° latitude north towards Asia, closing the ancient Neotethyan Ocean (e.g. Burg 2011). This ocean plays a central role in reconstructing a number of orogenic systems extending from Europe to Asia (e.g. Şengör & Atayman 2009) (Fig. 1). However, debated palaeotectonic issues remain because exposures of the former ocean and its sub-basins are affected by plate collision, arc and terrane accretion, and later deformation. Problems that plague Neotethyan reconstructions include complicated structural relationships between sutures and amalgamated crustal fragments, unknown subduction polarity of oceanic arcs, unknown origin or tectonic setting of the oceanic fragments present in the collision zone, and the potential existence and locations of smaller, marginal ocean basins within the Neotethys itself (e.g. Şengör et al. 1984; Ustaömer & Robertson 1997; Robertson & Collins 2002; Aitchison et al. 2007; Şengör & Atayman 2009). Exposed remnants of Neotethyan ophiolites are present across the Himalayas and have the potential to provide considerable insight into the timing and processes involved during Indo-Asia collision and the nature of Neotethyan ocean crust and upper mantle (e.g. Hébert *et al.* 2012; Hu *et al.* 2016).

The Spongtang ophiolite (Figs 1 and 2) was described as early as the late 19th century (Lydekker 1880, 1883; La Touche 1888; MacMohan 1901) and has been the focus of several detailed structural, geochemical and geochronological studies since the 1980s (Bassoullet et al. 1980; Andrews-Speed & Brookfield 1982; Reibel & Reuber 1982; Kelemen & Sonnenfeld 1983; Reuber 1986a, b; Reuber & Colchen 1986; Reuber et al. 1987; Niorthe & Reuber 1988; Searle et al. 1997; Corfield et al. 1999, 2001, 2005; Corfield & Searle 2000; Pedersen et al. 2001; Mahéo et al. 2004; Baxter et al. 2010). The ophiolite is one of several isolated basicultrabasic massifs exposed across the collision zone (Fig. 1) (Brookfield 1977; Reuber 1986a, b; Reuber et al. 1983, 1987, 2005; Hébert et al. 2012; Xu et al. 2015). A comparison of the Spongtang ophiolite complex with others along the Himalayan arc and elsewhere along the site of Neotethyan ocean closure is challenged by limited and problematic age data and alteration



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Fig. 1. (a) Location of ophiolites across the eastern Neotethyan domain (after Reuber 1990). Box indicates the location of the map in (b). C-R arc, Chagai-Raskoh arc, K, Kandahar arc. (b) Simplified tectonic map showing the locations of ophiolites across the Himalayan arc after Hébert et al. (2012). SSZ, Shyok Suture Zone; ZSZ, Zanskar Suture Zone; YZSZ, Yarlung Zangbo Suture Zone. Himalayan fold-thrust belt structures: MCT, Main Central Thrust, STDS, South Tibet Detachment System, MBT, Main Boundary Thrust; MFT, Main Frontal Thrust; MMT, Main Mantle Thrust. Box indicates the location of the Spongtang ophiolite and Spong arc (Fig. 2). We also show locations of the Dras arc and Nidar ophiolite.

owing to the influence of multiple tectonic events (Ghose et al. 2014).

In this study, ophiolite samples collected by Ingrid Reuber in 1986 (Table 1; Fig. 2) were petrographically and geochemically analysed and dated. Ages were obtained from various units of the ophiolite section via *in situ* (in thin section) secondary ion mass spectrometry (SIMS) (U–Pb zircon) and whole-rock noble gas spectrometry 40 Ar/ 39 Ar geochronology to determine its crystallization and deformation history. We apply advances in the interpretation of pyroxene rare earth element (REE) distributions and conventional major-element based geothermometry to discern a cooling rate over high-temperature cooling intervals, an approach

that may allow interpretations to be made regarding its geological evolution (Liang *et al.* 2013; Dygert & Liang 2015).

Geological background

The Spongtang ophiolite is a tectonic thrust slice that overlies Permian to Paleocene–Eocene sediments of the northern Indian passive margin (Andrews-Speed & Brookfield 1982; Ahmad *et al.* 2008; Green *et al.* 2008; van Hinsbergen *et al.* 2012; Clift *et al.* 2014; Gibbons *et al.* 2015) (Figs 1 and 2). It is characterized by a complete ophiolite stratigraphy, dominated by peridotites (Reuber *et al.* 1989; Corfield *et al.* 2001; Ghose *et al.* 2014), and is divided



Fig. 2. (a) Geological map of the Spongtang ophiolite and Spong arc. Sample locations (SPO-#) after Reuber *et al.* (1989) and Corfield *et al.* (2001). Map source: Esri, DigitalGlobe, GeoEye, Earthstar Geographics, CNES/Airbus DS, USDA, USGS, AeroGRID, IGN and the GIS User Community. Ages after Reuber *et al.* (1989), Pedersen *et al.* (2001) and Clift *et al.* (2014). (b) Cross-sections across A–A', B–B' and C–C' after Corfield *et al.* (2001). Section A–A' shows sample locations of SPO4, SPO20, SPO22, SPO25 and SPO26. Section B–B' shows sample locations of SPO19 and SPO17. Section C–C' shows sample locations of SPO10, SPO14 and SPO15. (See Table 1 for rock descriptions.)

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Fig. 3. Plot summarizing available age data (N=number) from the Spongtang ophiolite, Spong arc and associated sediments, including the timing of significant events. Age data after Reuber *et al.* (1989, 1990), Pedersen *et al.* (2001), Mahéo *et al.* (2004) and Hébert *et al.* (2012). Interpretation of timing of rifting after Reuber *et al.* (1987). The age for the Spongtang ophiolite generation is after Pedersen *et al.* (2001). Reuber *et al.* (1989) suggested a major intra-oceanic shearing event based on the K–Ar age of sample SPO26. The activity of the Spong arc is based on multiple zircon ages in the dashed box labelled 'volcanics'. Initiation of subduction magmatism was suggested by Mahéo *et al.* (2004) based on ⁴⁰Arl³⁹Ar ages. Spongtang ophiolite obduction may have occurred in the Late Cretaceous or coeval with Indo-Asia collision (for discussion, see Ghose *et al.* 2014). Re-thrusting of the ophiolite section may have occurred post-Eocene (Corfield *et al.* 1999, 2001). Ages from the Stumpata formation are from zircons dated by Clift *et al.* (2014), and chert biostratigraphy after Baxter *et al.* (2010). Some sample numbers are labelled (SPO-#, this study; SP-#, Mahéo *et al.* 2004).

by the Photang valley into a western area with mantle rocks and an eastern region of exposed crustal assemblages. Northwest of the Photang valley, the lowermost Photang thrust sheet consists of distinct alkaline volcanic units, overlain by Permian to Cretaceous limestones, which most probably were seamounts in the Neotethyan Ocean (Corfield et al. 1999). Above are two ultrabasic thrust sheets of suboceanic mantle rocks whose contact Reuber (1986a) considered an intra-oceanic thrust. South of the Photang valley, a thin mélange with limestone exotic rocks underlies an upper ophiolitic thrust sheet with gabbros, sheared dykes and pillow lavas overlain by Late Cretaceous andesites interbedded with latest Valanginian to early to mid-Aptian radiolarian cherts (Corfield et al. 1999; Baxter et al. 2010). The Spongtang massif can thus be divided into two distinct suites of related affinities. The older suite is the Jurassic Spongtang ophiolite, comprising two (stratigraphically lower and upper) mantle units as well as crustal rocks, which was deformed in the early Cretaceous. The younger suite is the overlying Spong arc, a c. 500 m thick, volcano-sedimentary package that developed in an island-arc environment (Reuber et al. 1989; Corfield et al. 2001). The arc is similar in age to the Dras arc along the Yarlung Zangbo Suture Zone to the north (Fig. 1) (see review by Hébert et al. 2012).

The timing of formation and emplacement of the ophiolite and its oceanic island arc cover (Spong arc) and its eventual collision with northern India are controversial but essential for deciphering the history of plate convergence during the closing of the Neotethyan ocean (Rolland *et al.* 2002; Clift *et al.* 2014; Hu *et al.* 2016). Existing age constraints are based on direct dating of rocks

associated with the ophiolite, as well as dating of overlying and underlying sediments (Corfield *et al.* 2001; Baxter *et al.* 2010; Clift *et al.* 2014). We summarize available radiometric age constraints in Figure 3. Ages from minerals extracted from the ophiolite sequences, in general, yield large uncertainties reflecting both analytical issues (i.e. excess argon in K–Ar age and complex 40 Ar/ 39 Ar gas release patterns) and the unit's crystallization and deformation history.

Reuber et al. (1989, 1990) reported K-Ar amphibole ages from some of the same rocks analysed here that range from 231 ± 23 to 124 ± 75 Ma. All have large uncertainty (the latter age is not plotted in Fig. 3) and are attributed to oceanic deformation and accretion, with some affected by excess argon. Sample SPO26, an amphibolite from the Spongtang metamorphic sole (Fig. 2), yields 135 ± 4 Ma, suggested to time intra-oceanic thrusting (Reuber et al. 1990). The most commonly reported crystallization age for the Spongtang ophiolite is 177 ± 1 Ma, based on three fractions of zircons separated from a diorite intruding high-level gabbros of the ophiolite (Pedersen et al. 2001) (see Fig. 2 for sample location). The timing of activity in the overlying Spong arc is based on a series of zircon ages extracted from andesite that average 88 ± 5 Ma, although zircons from other andesite samples are older (Pedersen et al. 2001) (Fig. 3). The Early Jurassic zircon crystallization age makes the Spongtang ophiolite the oldest known Neotethyan ophiolite (Ghose et al. 2014).

Metamorphic amphiboles from Spongtang diorite dykes intruding lower peridotites yield disturbed ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ age spectra interpreted to record the initiation of subduction magmatism at *c*.

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Table 1. Ophiolite samples analysed in this study

Sample*	Reuber sample†	Sample description: detailed summary of mineralogy, textures, previously reported geochemistry and K-Ar ages‡
Crustal ophiolite sequence		
SPO17§	L130	Basalt subvolcanic: blocky plagioclase; interstitial clinopyroxene; serpentinite, matrix and alteration minerals in veins
SPO19§	P477	Basalt pillow lava: effectively serpentinite; calcite and quartz cross-cutting veins, altered olivine, some intact plagioclase laths
Mantle ophiolite sequence, upper unit		
SPO4§¶	LK51	Diorite dyke, K–Ar 139 ± 8 Ma: plagioclase altered to sericite with some intact grains, twinned clinopyroxene, chlorite present
SPO6§¶	LK100, 89/127	Diorite dyke, K-Ar 169.0 ± 9.5 Ma: reported foliated, green amphibole, 'pegmatitic' textures, coarsest grained of all samples
SPO10§¶	P117	Plagiogranite: rare in region, undulatory quartz, some with chequerboard textures, dominated by grain boundary migration
SPO14§	L217	Isolated dyke: plagioclase laths with interstitial clinopyroxene, blue chlorite, rare myrmekite
SPO15§	P102C	Isolated dyke: amphibole analysis reported; extremely altered assemblage dominated by sericite and quartz + few opaque mineral grains
SPO20¶	L103	Deformed diabase dyke: clear foliation defined by clinopyroxene, sericite-plagioclase blades and quartz with grain boundary migration textures
SPO22§	L106, 81/103	Deformed diabase dyke, K–Ar 231 ± 23 Ma: deformed owing to intra-oceanic thrusting, foliated, green amphibole; pyroxene or amphibole in a matrix of altered plagioclase; chlorite veins
Mantle ophiolite sequence, lower unit		
SPO25**	LE114A	Harzburgite: >75% olivine, <25% orthopyroxene, minor spinel and clinopyroxene; foliated
SPO26**	P338	$Metamorphic \ sole \ (amphibolite), K-Ar \ 135 \pm 4 \ Ma: \ green \ hornblende + plagioclase + epidote; \ foliated, \ secondary \ quartz \ veins$

*Sample locations are shown in Figure 2.

*Sample number as designated by Reuber et al. (1989, 1990).

\$Summary of descriptions and available K-Ar of amphibole provided by Reuber et al. (1989, 1990). Amphibole analysis of SPO15 by Reuber et al. (1990).

§⁴⁰Ar/³⁹Ar whole-rock ages obtained in this study.

¶U-Pb zircon ages obtained via in situ SIMS in this study.

**Mineral compositions obtained in this study.

130 Ma (Mahéo *et al.* 2004). None yield plateau ages as defined by Ludwig (2008) and the results demonstrate the problems associated with obtaining reliable 40 Ar/ 39 Ar data from the amphibole separates from these rocks, as they are affected by excess argon. The results also suggest the potential presence of alteration minerals recorded by Ca/K as the grains degassed during the step-heating process. Our new argon geochronology data show similar problems and are discussed in detail in the results section below.

Regarding its tectonic affinity, the Spongtang ophiolite has been proposed to have formed at a back-arc basin (Reuber et al. 1983), a slow spreading centre near a transform fault (e.g. Reuber 1990) or a mid-ocean ridge-like setting with an normal mid-ocean ridge basalt (N-MORB) mantle source previously enriched in large ion lithophile elements (LILE) by a metasomatic event that occurred in a suprasubduction zone (SSZ) (Pedersen et al. 2001; Mahéo et al. 2004). Several models have been proposed regarding its emplacement history. Assuming a tectonic scenario similar to that for the Samail ophiolite in Oman (Brookfield 1977; Searle et al. 1988), the Spongtang ophiolite may have been initially emplaced in the Late Cretaceous onto the northern Indian passive margin and deformed in the Late Eocene during the final closure of the Neotethyan Ocean (Brookfield & Reynolds 1981; Corfield et al. 1999; model B of Clift et al. 2014). Alternatively, obduction may have been entirely a late Eocene event (Colchen et al. 1987; Mahéo et al. 2004; Fuchs 2005; model A of Clift et al. 2014). Ophiolitic detritus in underlying sedimentary units suggests that the Spongtang ophiolite was situated on or near the northern Indian margin by the Paleocene (Garzanti et al. 1987; Searle et al. 1990). The detritus is also found in the Paleocene deep-water sediments of the north Indian margin in southern Tibet (Ding et al. 2005). A significant presence of Mesozoic detrital zircons $(129 \pm 7 \text{ Ma})$ unrelated to the unit indicates that if obduction occurred on the Indian margin during the collision, the Spongtang ophiolite was submarine or un-eroded until later re-thrusting, or emplacement occurred after c. 61 Ma (Clift *et al.* 2014). This paper presents additional geochronological and geochemical data from selected units of the Spongtang ophiolite to evaluate proposed episodes of its tectonic history and place constraints on the timing and dynamics of its crystallization and deformation.

Materials and methods

Samples and petrography

This study focuses on the analysis of 11 rocks collected from the Spongtang ophiolite (Table 1; Fig. 2). All samples are deformed and altered diabase or diorite dykes, except a plagiogranite (SPO10), harzburgite (SPO25), amphibolite (SPO26) and two subvolcanic assemblages (SPO17 and SPO19). Although no metamorphic sole is purported to exist below the Spongtang ophiolite (Pedersen *et al.* 2001), we consider sample SPO26 to be a component of the sole based on its structural location, petrography and mineralogy. All rocks were examined initially with a petrographic microscope to document mineralogy and textures. As indicated in Table 1, selected samples were subjected to geochemical (mineral major and trace element) and geochronological (40 Ar/³⁹Ar whole-rock and U–Pb zircon SIMS) analysis.

Mineral chemistry of samples SPO25 and SPO26

Major elements in clinopyroxene (n=3 grains, 19 spots), orthopyroxene (n=3 grains, 31 spots), spinel (n=2 grains, 19 spots) and olivine (n=2 grains, 16 spots) in harzburgite SPO25 and amphibole in amphibolite sample SPO26 (n=3 grains, n=9 spots)

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Fig. 4. Petrographic images in cross-polarized light of crustal ophiolite sequences: (a) SPO17, two images; (b) SPO19. Mineral abbreviations after Siivola & Schmid (2007). (See Fig. 2 for sample locations.)

were determined using the JEOL JXA-8200 electron microprobe at the University of Texas at Austin. Detailed results are presented in the supplementary material. Analyses were conducted during the same analytical session using standard conditions (a focused beam, 20 kV accelerating voltage, c. 20 nA beam current) and a 10element procedure (SiO₂, TiO₂, Al₂O₃, Cr₂O₃, FeO, MnO, MgO, Na₂O, K₂O, CaO). Natural and synthetic standards were processed using a JEOL ZAF matrix correction procedure, and results were evaluated to ensure they fit mineral stoichiometry.

Li, Na, Mg, Al, V, Sr, Pb, Y, high field strength elements (HFSE) and REE were analysed in orthopyroxene (n = 10 spots) and clinopyroxene (n = 8 spots) in sample SPO25 and amphibole in SPO26 (n = 6 spots) by laser ablation inductively coupled plasma mass spectrometry (LA-ICP) using a New Wave UP193FX laser with a dual-volume sample chamber coupled to an Agilent 7500ce quadrupole ICP-MS system. We applied a laser fluence of c. 3 J cm^{-2} in one analytical session at a frequency of 10 Hz with 60 s ablations that penetrated through the samples completely. The process ensured the introduction of measurable quantities of REE into the ICP-MS system. Synthetic NIST glasses (610, 612 and 614) (Jochum et al. 2011) and a natural basalt (BCR-2) (Chen et al. 2000) were used as standards. Data were processed in LASY BOY (Sparks 2011) with Si as the internal standard. Spot sizes ranged from 40 to 120 µm, using the largest spots possible while avoiding cracks or alteration zones. Standards were analysed in triplicate before and after the unknowns using the same spot sizes as the unknowns. Detailed results are presented in the supplementary material.

Argon geochronology

Eight samples (SPO4, SPO6, SPO10, SPO14, SPO15, SPO17, SPO19 and SPO22) were subjected to ⁴⁰Ar/³⁹Ar whole-rock dating at the Indian Institute of Technology Laboratory at Mumbai. Portions of these rocks were crushed in a tungsten mill, and c. 200 mg powder (120-180 µm size fraction) was cleaned with 1% HCl and deionized water before packing in aluminium capsules for irradiation. The samples were irradiated with a flux monitor (Bern 4 Biotite, 17.3 ± 0.2 Ma, Flisch 1982; 17.21 Ma, Baksi et al. 1996) in the DHRUVA reactor at Bhabha Atomic Research Centre, Mumbai, India in one batch for c. 120 h. Pure CaF₂ and K₂SO₄ salts were also irradiated with the samples to estimate the Ar produced by nuclear reactions on Ca and K isotopes. The following correction factors are used for correcting the reactor produced interfering Ar isotopes: $({}^{36}\text{Ar}/{}^{37}\text{Ar})_{Ca} = 0.002227, ({}^{39}\text{Ar}/{}^{37}\text{Ar})_{Ca} =$ 0.034342 and $({}^{40}\text{Ar}/{}^{39}\text{Ar})_{\text{K}} = 0.0381106$. ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ ages were calculated using the constants recommended by Steiger & Jäger (1977); use of Renne et al. (2010, 2011) constants would increase the age (in the age range of this study) by c. 0.9% (see discussions

by Renne *et al.* 2011 and Schwarz *et al.* 2011), indistinguishable within their state uncertainties. The relative variation in the neutron fluence within the capsule was monitored by pure Ni wire that was also irradiated with the samples. Argon gas was extracted via step heating in a furnace, starting at 450°C with intervals from 25 to 50° C going up to a maximum of 1350°C, except for sample SPO19, which experienced a maximum of 1050°C. The number of steps varied from 14 to 20. The gas released at each step was cleaned using Ti–Zr getters and analysed for argon isotope ratios in the ARGUS VI multi-collector mass spectrometer (ThermoFisher) at the National Facility at Indian Institute of Technology (IIT), Mumbai, India. Detailed results are presented in the supplementary material.

U-Pb zircon geochronology

All samples were examined using a scanning electron microscope (SEM) for zircon grains for in situ SIMS analysis. Zircons were found only in samples SPO4, SPO6, SPO10, SPO20 and SPO26. The grains ranged from $10-15 \,\mu\text{m}$ to c. $100 \,\mu\text{m}$ in length. Their locations were documented, and portions of the thin sections containing the grains were cut from the samples using a low-speed saw. The thin section chips were mounted in epoxy with blocks of zircon age standard AS3 (1099.1 \pm 0.5 Ma, Schmitz et al. 2003), coated in gold, and analysed using the UCLA CAMECA ims 1270-HR ion microprobe (sample SPO4 only) and CAMECA ims 1280-HR ion microprobe at Heidelberg University (all other samples). In both laboratories, a calibration curve was developed using U and Pb isotopic data sputtered from the age standards. For sample SPO4, 23 spots were placed on standard AS3, which created a calibration curve in UO⁺/U⁺ = $5.385 \pm 0.071 - 0.468 \times (Pb/U)$, relative sensitivity factor), reproducing the standard age to 1093 ± 40 Ma. For all of the other samples, 16 spots were placed on standard AS3, which created a calibration curve in $UO_2^+/U^+ = 1.434 \pm 0.014 - 0.285 \times$ (Pb/U, relative sensitivity factor), which reproduced the standard age to 1100 ± 37 Ma. After dating, all zircons were imaged using backscattered electrons (BSE), secondary electrons (SE) and cathodoluminescence (CL) to determine zoning characteristics and location of the ion beam.

Results

Petrography

We show images of samples analysed in this study in Figures 4–6, which illustrate their mineralogy, alteration and deformation textures. All rocks are affected by varying degrees of alteration, expected as they experienced a complex crystallization and deformation history, but most preserve some degree of primary igneous mineralogy and textures. We compare our petrographic



Fig. 5. Petrographic images in cross-polarized light of upper mantle unit ophiolite samples: (a) SPO4; (b) SPO6; (c) SPO14; (d) SPO15; (e) SPO10, two images; (f) SPO20; (g) SPO22. Mineral abbreviations after Siivola & Schmid (2007). (See Fig. 2 for sample locations.)



Fig. 6. Petrographic images in cross-polarized light of lower mantle unit ophiolite samples: (a) SPO25, two images; (b) SPO26. Mineral abbreviations after Siivola & Schmid (2007). (See Fig. 2 for sample locations.)



Fig. 7. BSE images of samples: (**a**–**c**) SPO25; (**d**) SPO26. Major minerals are labelled; abbreviations after Siivola & Schmid (2007). Both the electron microprobe and laser ablation spots (scaled for analytical size) are indicated. Spot numbers are labelled. (See supplementary material for details regarding the results from each spot.) In (**a**) and (**c**), we outline the approximate grain boundaries for orthopyroxene grains. (**e**)–(**h**) show the results of compositional transects identified in the BSE images. (**e**) shows olivine Fo (= (Mg/(Mg + Fe)) × 100) and orthopyroxene En (= (Mg/(Mg + Fe + Ca)) × 100); (**f**) shows spinel Cr/(Cr + Al); (**g**) plots mole fraction Wo (= Ca/ (Ca + Mg + Fe_{total})); (**h**) indicates the amphibole Fe (atoms per formula unit). Amphibole totals were estimated assuming 23 oxygens and all Fe is present as Fe²⁺. Transects are identified. (**h**) also includes the approximate locations where compositions of points 46, 47 (core) and 48 (rim) analyses are located.

	SPO25: Olivine average all $n = 16 \ (\pm 1\sigma)^*$	SPO25: Clinopyroxene average all $n = 19 \ (\pm 1\sigma)$	SPO26: Amphibole average all $n = 9$ (±1 σ)	SPO26: Amphibole bright BSE $n = 5 \ (\pm 1\sigma)$	SPO26: Amphibole darker BSE $n = 4 \ (\pm 1\sigma)$
SiO ₂	39.8 (0.4)	51.5 (0.7)	37.2 (0.6)	37.1 (0.38)	37.4 (0.77)
TiO_2	0.06 (0.5)	0.19 (0.07)	0.20 (0.13)	0.29 (0.09)	0.09 (0.05)
Al_2O_3	-†	3.24 (0.48)	24.0 (1.1)	23.4 (0.91)	24.8 (0.78)
Cr_2O_3	0.01 (0.02)	0.79 (0.19)	_	_	_
FeO	9.51 (0.23)	2.38 (0.17)	11.6 (1.3)	12.4 (1.0)	10.5 (0.6)
MnO	0.15 (0.02)	0.09 (0.02)	0.14 (0.04)	0.13 (0.03)	0.15 (0.05)
MgO	50.1 (0.5)	17.3 (0.3)	0.05 (0.03)	0.07 (0.03)	0.04 (0.01)
CaO	0.01 (0.01)	24.6 (0.4)	24.2 (0.3)	24.1 (0.2)	24.4 (0.3)
Na ₂ O	-	0.07 (0.01)	_		
Total	99.6 (0.7)	100.2 (0.6)	97.5 (1.01)	97.5 (0.8)	97.4 (1.4)
Si	0.975 (0.01)	1.865 (0.017)	5.597 (0.040)	5.601 (0.041)	5.591 (0.043)
Ti	-	0.005 (0.002)	0.023 (0.014)	0.033 (0.010)	0.010 (0.005)
Al	-	0.138 (0.021)	4.249 (0.159)	4.155 (0.136)	4.367 (0.099)
Cr	-	0.023 (0.005)	_	_	-
Fe	0.194 (0.004)	0.072 (0.005)	1.458 (0.172)	1.570 (0.134)	1.317 (0.089)
Mn	-	0.003 (0.001)	0.018 (0.005)	0.017 (0.004)	0.019 (0.006)
Mg	1.827 (0.01)	0.936 (0.017)	0.012 (0.006)	0.015 (0.007)	0.009 (0.002)
Ca	-	0.953 (0.013)	3.899 (0.028)	3.897 (0.027)	3.903 (0.032)
Na	-	0.005 (0.001)	_	_	-
Fo‡	90.2 (0.2)				
Fa	9.61 (0.21)				
Te	0.16 (0.01)				
Wo		0.486 (0.007)			
En		0.477 (0.006)			
Fs		0.037 (0.003)			
ХСа			0.727 (0.023)	0.711 (0.016)	0.747 (0.013)
XFe			0.271 (0.023)	0.286 (0.017)	0.252 (0.013)

Table 2. Compositions of olivine, clinopyroxene (sample SPO25) and amphibole (SPO26)

**n* is number of analyses; $(\pm 1\sigma)$ is the standard deviation of the value of the analyses. Olivine atoms per formula unit (a.p.f.u.) based on three cations and four oxygens. Clinopyroxene a.p.f.u. based on four cations and six oxygens. Amphibole analyses based on 23 oxygens and two hydrogens.

†-, measured but not detected.

observations with those of the same samples by Reuber *et al.* (1983, 1990, 2005).

Samples from the upper crustal ophiolite sequence include a subvolcanic basalt (SPO17) and pillow lava (SPO19) (Fig. 4). The subvolcanic basalt assemblage (Fig. 4a and b) consists primarily of blocky plagioclase that is variably altered and aligned to define the rock's foliation. Clinopyroxene and opaque minerals fill the interstitial spaces between larger plagioclase grains. This sample has cross-cutting veins filled with similar mineralogy to the rock itself, but with smaller grain sizes. The pillow lava sample (SPO19) is extremely altered with only relict plagioclase laths in a serpentinite–chlorite–sericite matrix (Fig. 4c). The rock is cut by veins with calcite cores surrounded by quartz candles with *c*-axes that align towards the core. The edges of some veins spread diffusively into the altered rock matrix.

Petrographic images of samples from the upper mantle unit ophiolite sequences are shown in Figure 5. Samples SPO4 (K–Ar amphibole age of 139 ± 8 Ma) and SPO6 (K–Ar amphibole age of 169.0 ± 9.5 Ma; see Reuber *et al.* 1989, 1990) are from the pegmatite dykes described by Reuber *et al.* (1989) and consist primarily of large clinopyroxene grains with variable altered plagioclase grains. Zircon grains in SPO4 are rare and are found primarily on the edges of clinopyroxene. These dyke samples have the coarsest grain size and thus differ from samples SPO14 and SPO15, which have similar mineralogy but are composed of smaller grain sizes. Sample SPO15 is more altered than sample SPO14 and is dominated by sericite matrix with quartz and some opaque minerals. We highlight a quartz vein in sample SPO15, which has finer-grained and coarser-grained sections (Fig. 5d). Sample SPO10 is a plagiogranite dominated by quartz-plagioclase mineralogy with minor opaque minerals and sericite owing to the alteration of plagioclase. The majority of the rock has larger quartz grains with lobate grain boundaries and chequerboard textures, but anastomosing veins that cut the sample show much smaller grain sizes. SPO10 also contains minor amounts of frayed biotite that has largely altered to chlorite. These biotite grains are where the majority of zircon in the rock is present.

SPO20 and SPO22 are both deformed diabase dykes, with a clear foliation defined by clinopyroxene, sericite–plagioclase blades and quartz with grain boundary migration textures. Sample SPO20 has quartz veins parallel to the foliation that show undulatory extinction and grain-boundary migration textures. In sample SPO22, we find cross-cutting veins with asymmetrical mineral assemblages, with chlorite–sericite grains opposite coarser plagioclase and quartz. Reuber *et al.* (1989) suggested that the deformation textures observed in the deformed diabase dykes were produced during intraoceanic thrusting.

Petrographic images of lower mantle unit samples SPO25 and SPO26 are shown in Figure 6. Sample SPO25 is a harzburgite and contains olivine (>75%) with lesser amounts of orthopyroxene, clinopyroxene and spinel. The rock is altered by a web-like array of serpentine and chlorite veins that fill grain boundaries and transect the rock. Amphibolite sample SPO26 (K–Ar amphibole, 135 ± 4 Ma, Reuber *et al.* 1989) originates from the Spongtang ophiolite's metamorphic sole. The rock is composed primarily of hornblende and epidote and is foliated. These rocks were the focus for further geochemical analyses described in the next section.

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Table 3. Compositions of spinel and orthopyroxene in sample SPO25

	Spinel average all $n = 19 \ (\pm 1\sigma)^*$	Spinel average core $n = 12 \ (\pm 1\sigma)$	Spinel average rim $n = 2 \ (\pm 1\sigma)$	Opx average all $n = 29 \ (\pm 1\sigma)$	Opx average lower En, $n = 5 \ (\pm 1\sigma)$	Opx average higher En, $n = 24 \ (\pm 1\sigma)$
SiO ₂	0.07 (0.03)	0.06 (0.01)	0.13 (0.01)	54.4 (0.9)	54.6 (0.1)	54.4 (1.0)
TiO ₂	0.14 (0.05)	0.14 (0.05)	0.18 (<0.01)	0.07 (0.03)	0.09 (0.03)	0.06 (0.03)
Al_2O_3	42.3 (1.3)	41.5 (0.3)	45.1 (0.5)	3.24 (0.25)	3.52 (0.13)	3.19 (0.24)
Cr_2O_3	24.8 (1.3)	25.5 (0.3)	22.1 (0.4)	0.59 (0.12)	0.70 (0.08)	0.57 (0.12)
FeO	15.3 (0.3)	15.4 (0.1)	14.4 (0.2)	6.26 (0.15)	6.02 (0.08)	6.31 (0.11)
MnO	0.20 (0.02)	0.20 (0.02)	0.20 (0.03)	0.15 (0.02)	0.16 (0.01)	0.15 (0.02)
MgO	17.3 (0.2)	17.2 (0.1)	17.8 (0.01)	33.9 (0.5)	32.8 (0.2)	34.1 (0.3)
CaO	-†	_	_	1.04 (0.62)	2.37 (0.34)	0.78 (0.17)
Total	100.0 (0.4)	100.0 (0.4)	100.0 (0.01)	99.7 (0.9)	100.2 (0.3)	99.6 (1.0)
Si	0.002 (0.001)	0.002 (<0.001)	0.004 (<0.001)	1.882 (0.017)	1.883 (0.005)	1.881 (0.019)
Ti	0.003 (0.001)	0.003 (0.001)	0.004 (<0.001)	0.001 (0.001)	0.002 (0.001)	0.001 (0.001)
Al	1.379 (0.033)	1.359 (0.007)	1.454 (0.015)	0.132 (0.010)	0.143 (0.005)	0.130 (0.009)
Cr	0.544 (0.030)	0.561 (0.006)	0.479 (0.010)	0.016 (0.003)	0.019 (0.002)	0.016 (0.003)
Fe ³⁺	0.067 (0.007)	0.071 (0.005)	0.053 (0.006)	-	-	-
Fe ²⁺	0.287 (0.006)	0.287 (0.004)	0.277 (0.001)	0.176 (0.032)	0.174 (0.002)	0.176 (0.035)
Mn	0.005 (0.001)	0.005 (<0.001)	0.004 (0.001)	0.005 (<0.001)	0.005 (<0.001)	0.005 (<0.001)
Mg	0.713 (0.007)	0.713 (0.004)	0.726 (0.002)	1.744 (0.031)	1.687 (0.013)	1.755 (0.020)
Ca	-	-	-	0.038 (0.023)	0.088 (0.012)	0.029 (0.006)
Cr#‡	0.283 (0.01)	0.292 (0.003)	0.248 (0.006)			
En	. ,	× ,		0.888 (0.010)	0.866 (0.010)	0.892 (0.003)
Fs				0.092 (0.002)	0.089 (0.002)	0.093 (0.001)
Wo				0.020 (0.012)	0.045 (0.012)	0.015 (0.012)

*n is number of analyses; $(\pm 1\sigma)$ is the standard deviation of the value of the analyses. Spinel atoms per formula unit (a.p.f.u.) based on three cations and four oxygens. Orthopyroxene a.p.f.u. based on four cations and six oxygens.

†-, measured but not detected.

Abbreviations: Cr#, (spinel Cr/(Cr + Al)), Wo (orthopyroxene Ca/(Ca + Fe + Mg)), En (orthopyroxene, Mg/(Ca + Fe + Mg)), Fs (orthopyroxene, Fe/(Ca + Fe + Mg)).

Mineral chemistry of samples SPO25 and SPO26

Figure 7 shows the locations and compositions across transects of olivine, spinel, orthopyroxene and clinopyroxene in sample SPO25 and amphibole in SPO26. Our analyses of olivine and clinopyroxene grains in sample SPO25 indicate lack of zoning (Table 2).

Olivine forsterite contents (Fo, $(Mg/(Mg + Fe_T)) \times 100)$ average 90.2 ± 0.2 (n = 16), and clinopyroxene enstatite compositions (En, Mg/(Mg + Ca + Fe_T)) average 0.48 ± 0.01 (n = 19). We find that most SPO25 orthopyroxene En contents average 0.892 ± 0.002 (n = 26), with five analyses that are outside this value (from 0.877 to 0.860) where the electron beam excited clinopyroxene exsolution



Fig. 8. (a) Plot of olivine forsterite number (Fo) v. spinel chromium number (Cr#) after Arai (1994). OSMA, olivine–spinel mantle array; FMM, fertile MORB mantle. The 5 and 10 kbar curves are from Sobolev & Batanova (1995); the 15 kbar curve is from Jacques & Green (1980) (see Pirard *et al.* 2013). Melting trend (annotated by % melting) is from Arai (1994). (b) Plot of spinel Mg# v. Cr#. Abyssal and SSZ fields are the same as in (a). Partial melting curve after Hirose & Kawamoto (1995) and Whattam *et al.* (2011). (c) Plot of TiO₂ wt% v. Cr# of spinel. Partial melting curve and abyssal basalt composition after Choi *et al.* (2008). (d) Melt depletion of Spongtang sample SPO25, based on Cr# of spinel after the Hellebrand *et al.* (2001) relationship $F = 10\ln(Cr#) + 24$, where *F* is melt fraction. This relationship is valid for spinel Cr# values between 0.1 and 0.6. In (a)–(c), abyssal peridotites are after Warren (2016) and SSZ ophiolites after Ishii *et al.* (1992). Included in each panel are data from sample SPO25 (red rectangles) and peridotites from the upper and lower sections of the Spongtang ophiolite from Mahéo *et al.* (2004) (green and black ovals; those with Cr# >0.6 are dunites). Averages from SPO25 cores and rims are also shown with ranges indicated using error bars.

lamellae (Table 3). The compositions of these lamellae grains fit the stoichiometry of pyroxene and have oxide wt% totals of 99–100 wt % (Table 3). The Mg# (Mg/(Mg + Fe_T)) of spinel in sample SPO25 increases from core to rim (from 0.666 ± 0.001 to 0.687 ± 0.001), whereas spinel Cr# (Cr/(Cr + Al)) decreases (from 0.292 ± 0.003 to 0.248 ± 0.015) (Fig. 7, Table 3). BSE images of amphibole in sample SPO26 suggest that the mineral is zoned, and its brighter cores have slightly higher Fe contents and lower Ca contents compared with rims (Table 2).

In plots of olivine and spinel mineral chemistry, sample SPO25 falls directly within the compositional field defined by abyssal peridotites (Fig. 8a–c). According to methods presented by Hellebrand *et al.* (2001), the samples appear to have experienced $11.4 \pm 0.6\%$ melt extraction (Fig. 8d) assuming near-fractional



melting of a 'fertile MORB mantle' source. Although data from this sample overlap those for harzburgites and lherzolites reported by Mahéo *et al.* (2004) for upper and lower levels in the mantle section of the Spongtang ophiolite, spinels in Spongtang dunites have higher Cr contents consistent with interaction with a melt sourced from an SSZ setting.

Trace element data from clinopyroxene, orthopyroxene and amphibole grains are presented in Figure 9. For pyroxenes in general, the light REE (LREE) abundances for the minerals are generally below the instrumental detection limit, but middle REE (MREE) and heavy REE (HREE) abundances could be reliably quantified. Clinopyroxene exhibits REE concentrations 2.5–14 times higher than those for orthopyroxene. Extreme REE depletion characteristic of some SSZ-type settings (e.g. Dygert

> Fig. 9. (a) Incompatible minor elements in clinopyroxene; (b) incompatible minor elements in orthopyroxene. Samples that experienced larger extents of partial melting plot to the bottom left of the trends defined by abyssal peridotites (grey circles) and ophiolitic peridotites (blue circles). Ophiolite compositions are from the compilation of Dygert et al. (2017), and abyssal peridotite compositions are from PetDB and Warren (2016). Spongtang pyroxenes have compositions in the middle of the trends defined by ophiolites, suggesting moderate extents of partial melting. (c) Chondrite-normalized REE + Y concentrations in SPO25 clinopyroxene compared with abyssal and ophiolitic peridotites. (d) Chondritenormalized REE + Y concentrations in SPO26 amphibole, and SPO25 orthopyroxene and clinopyroxene. Continuous lines are averages, grey lines in the background show individual spot analyses and error bars show 1o standard deviations of replicate analyses.

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Fig. 10. (a) REE-in-two-pyroxene temperatures (T_{REE}) plotted against temperatures of the two-pyroxene solvus thermometer of Brey & Köhler (1990) (T_{BKN}) for abyssal peridotites (grey dots in peach field) and ophiolitic peridotites with mid-ocean ridge and SSZ tectonic affinities (blue and red circles respectively; compilation of Dygert & Liang 2015; Dygert *et al.* 2017). Samples that fall on the blue 1:1 line have T_{REE} and T_{BKN} in agreement, those left of the 1:1 line have higher T_{REE} than T_{BKN} . (b) Cooling curves (magenta lines) calculated using a modified form of the Dodson (1973) equation (Ganguly & Tirone 1999) overlain on data shown in (a). The path of a cooling sample in this temperature space is as follows: from an initial temperature on the blue 1:1 line, the sample moves to the left and then downward along a magenta cooling curve assuming a grain size of 0.5 mm. Initial temperatures are indicated to the right of the 1:1 line. Diffusion data are as assumed by Dygert & Liang (2015). The position of SPO25 in T_{BKN} - T_{REE} space indicates that it cooled at hundreds of degrees per million years or slower.

& Liang 2015) is not observed. Clinopyroxene and orthopyroxene have chondrite-normalized abundances and LREE depletion consistent with abyssal peridotites, suggesting adiabatic decompression melting of a MORB-mantle source beneath a spreading centre (Fig. 10).

Thermometry

Using REE, Y and and major element data, a REE-in-twopyroxene temperature (T_{REE}) was calculated from mineral compositions averaged from rim and core analyses using the approach described by Liang *et al.* (2013) and Dygert & Liang (2015) (Table 4). For comparison, temperatures from the twopyroxene solvus (Wells 1977; Brey & Köhler 1990) olivine–spinel (Fabriès 1979; Ballhaus *et al.* 1991) and Al-in-orthopyroxene (Witt-Eickschen & Seck 1991) thermometers were also calculated using the averaged mineral compositions for olivine, clinopyroxene, orthopyroxene, and core and rim compositions for spinel (Table 4).

Temperatures range from 656 to 1058° C depending on the thermometer. The olivine–spinel temperature decreases from core to rim by *c*. 15°C. The pyroxene solvus thermometers yield the lowest

temperatures with average Mg orthopyroxene compositions similar to those calculated if the highest values are used. We did not apply the low Mg contents to the barometer, as these are probably exsolution lamellae. The results of the pyroxene solvus thermometry are lower than the results estimated using olivine–spinel thermometry (by *c*. 17–185°C, depending on calibration). Thermometers based on trivalent element partitioning yield the highest temperatures (983–1058°C).

Figure 10 shows a comparison of the Spongtang T_{REE} and T_{BKN} with temperatures from abyssal peridotites and peridotites from ophiolites with mid-ocean ridge and SSZ tectonic affinities. SPO25 has a T_{REE} and T_{BKN} lower than any abyssal peridotite; SPO25 temperatures are consistent with the coldest ophiolitic peridotites, many of which have been interpreted to have SSZ tectonic affinities (see Dygert & Liang 2015).

⁴⁰Ar/³⁹Ar whole-rock geochronology

All whole-rock analyses yield disturbed ⁴⁰Ar/³⁹Ar age spectra and show older ages as step-heating progress (Figs 11–13). Groups of steps were linked based on trends identified in the step-heating ages and Ca/K ratios. In some cases steps belong to different groups, as

Table 4. Temperatures estimated for sample SPO25

Type*	Reference	T (°C) spinel core	T (°C) spinel rim	<i>T</i> (°C) Opx high Mg	T (°C) Opx average Mg
Olivine-spinel	Fabriès (1979)	841	817		
Olivine-spinel	Ballhaus et al. (1991)	817	794		
Pyroxene solvus	Brey & Köhler (1990)			656	658
Pyroxene solvus	Wells (1977)			774	800
Al-in-opx	Witt-Eickschen & Seck (1991)			983	990
REE-in-two-pyroxene†	Liang et al. (2013)			1058 ± 13	1027 ± 13

*Temperatures estimated assuming 10 kbar pressure if required for calibration.

†Temperature uncertainty determined from the temperature inversion (see supplementary material).



Fig. 11. Argon age spectra and Ca/K data v. cumulative ³⁹Ar fraction (% released) (**a**, **c**, **e**) and inverse isochron data (**b**, **d**, **f**) from dykes from the Spongtang ophiolite sequence. (**a**) and (**b**) are for SPO4; (**c**) and (**d**) are for SPO6; (**e**) and (**f**) are for SPO22. (See Fig. 2 for sample locations.) Shaded regions in the age spectra are steps used to generate the isochron ages in (**b**), (**d**) and (**f**). The red lines in (**b**), (**d**) and (**f**) are labelled based on the steps used to generate the inverse isochron data. We include some inverse isochron ages ($\pm 1\sigma$), initial ⁴⁰Ar/³⁶Ar, and mean square weighted deviation (MSWD) of the results. The age spectra include the average ages for groups of steps that are used to generate the inverse isochron ages. All errors are $\pm 2\sigma$. J factors are: SPO4, 0.00097891 \pm 0.0000047; SPO6, 0.000961587 \pm 0.000047; SPO22, 0.000897304 \pm 0.0000044.

the Ca/K data suggest that they may represent the degassing of potentially the same mineral. None of the samples have plateau ages as defined by Ludwig (2008). Inverse isochron plots show the influence of additional argon and ${}^{39}\text{Ar}/{}^{40}\text{Ar}$ values typically decrease with increasing step temperature. Sub-groups of these steps show linear correlations between ${}^{39}\text{Ar}/{}^{40}\text{Ar}$ v. ${}^{36}\text{Ar}/{}^{40}\text{Ar}$, which were used to determine some inverse isochron ages (e.g.

Korochantseva *et al.* 2007; Schwarz & Lippolt 2014). We attempted to identify trends in the inverse isochron results to decipher their relationship with plateau ages. However, we do not report results of isochrons that suggest a mixed age in Figures 11–13. All data are available in the supplementary material.

Overall, we find the youngest inverse isochron ages in the initial degassing steps (1–4 to 1–10) that range from 1.27 \pm



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Fig. 12. Argon age spectra and Ca/K data v. cumulative ³⁹Ar fraction (% released) (**a**, **c**, **e**) and inverse isochron data (**b**, **d**, **f**) from plagiogranite SPO10 (**a**, **b**) and isolated dykes from the Spongtang ophiolite sequence (SPO14 and SPO15, **c**–**f**). (See Fig. 2 for sample locations.) Shaded regions in the age spectra are steps used to generate the isochron ages in (**b**), (**d**) and (**f**). The red lines in (**b**), (**d**) and (**f**) are labelled based on the steps used to generate the inverse isochron ages $(\pm 1\sigma)$, initial ⁴⁰Ar/³⁶Ar, and MSWD of the results. J factors are: SPO10, 0.000985442 ± 0.0000048; SPO14, 0.000952118 ± 0.0000046; SPO15, 0.000903797 ± 0.0000044.

0.17 Ma ($\pm 2\sigma$, sample SPO15; Fig. 12) to 3.8 \pm 1.9 Ma (SPO6; Fig. 11). Corresponding average initial step ages from most samples are highly uncertain, but isolated dyke samples SPO14 and SPO15 yield 2.0 \pm 0.8 Ma and 2.3 \pm 0.6 Ma, respectively. The oldest of these initial step ages are Pliocene (5.1 \pm 1.9 Ma, sample SPO17; 5.5 \pm 3.6 Ma, SPO10). The results are probably related to the degassing of sericite or chlorite, minerals

associated with the alteration found in varying amounts in each sample (Figs 4–6).

After these initial steps, all samples trend towards older ages. The most precise Paleocene age is found in steps 10-14 of sample SPO14 (58.7 \pm 0.7 Ma), and more precise Paleocene to Late Cretaceous ages are found as in groups of steps, as opposed to inverse isochron ages. These results were generated in the final step-



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Fig. 13. Argon age spectra and Ca/K data v. cumulative ³⁹Ar fraction (% released) (**a** and **c**) and inverse isochron data (**b** and **d**) from basalt samples SPO17 (**a**, **b**) and SPO19 from the Spongtang ophiolite sequence (**c**, **d**). (See Fig. 2 for sample locations.) Shaded regions in the age spectra are steps used to generate the isochron ages in (**b**) and (**d**). The red lines in (**b**) and (**d**) are labelled based on the steps used to generate the inverse isochron data. We include some inverse isochron ages $(\pm 1\sigma)$, initial ⁴⁰Ar/³⁶Ar, and MSWD of the results. J factors are: SPO17, 0.000912299 \pm 0.0000044; SPO19, 0.000927222 \pm 0.0000045.

heating schedule and include 73.7 ± 2.2 Ma (sample SPO19, Fig. 13) and 101.7 ± 1.2 Ma (SPO15, Fig. 12). It should be noted that sample SPO17 had precise Cretaceous and Triassic ages for steps 11-14 (98.4 ± 2.3 Ma) and 14-16 (205 ± 3 Ma) (Fig. 13).

U-Pb zircon geochronology

Zircon grains were dated in samples SPO4, SPO6, SPO10 and SPO20 (Figs 14 and 15, Table 5). These grains are small (10–100 μ m in length). In sample SPO4, zircons are found at the edge of

clinopyroxene grains, whereas in sample SPO10, the grains are found in contact with and along the edges of frayed biotite. The grains are challenging to locate using the optics of the SIMS system and given the fine-grained nature of these rock types. We attempted to place the beam on the majority of the grain, but the ion microprobe spot overlapped the matrix in many cases during the dating process. Thus, we applied an aperture filter to measure isotopic data from only the radiogenic regions in the central region of the spot. After dating, the location of the spot was determined using SE. CL images of the grains provide additional information to assist with interpretations.



Fig. 14. Concordia diagrams for zircons in samples SPO4 and SPO6 (a), SPO10 (b) and SPO20 (c). Data from zircons reported by Pedersen *et al.* (2001) are shown in (b).

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Fig. 15. Images of zircons dated in samples SPO4 (a–d), SPO10 (e–i), SPO20 (j) and SPO6 (k–n). All panels are BSE images except (j), (l) and (n), which are CL images. Zircon grain number and $^{238}U^{-206}Pb$ ages (±1 σ) are indicated. The location of the ion microprobe spot is indicated. Insets in some panels show high-resolution CL images of the dated grains.

Overall, zircon ages from these samples range from Triassic $(234 \pm 18 \text{ Ma}, {}^{238}\text{U}/{}^{206}\text{Pb}$ age, $\pm 1\sigma$, sample SPO20) to Early Eocene $(42.4 \pm 0.5 \text{ Ma}, \text{ SPO6})$. Sample SPO20 yields the only Triassic result, whereas one zircon from sample SPO6 and six from SPO10 are Jurassic. All samples yield Cretaceous age zircons (SPO4 n = 4,

SPO6 n=2, SPO10 n=2 and SPO20 n=2). Only sample SPO6 contains one zircon that yielded Paleocene–Eocene ages.

The zircon that yields the Triassic age and the zircon with the second oldest age $(187.1 \pm 22.5 \text{ Ma}, \text{SPO6 zircon } 2)$ are discordant, whereas zircon 19b in sample SPO10 is reversely discordant

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Table 5. U-Pb age data from in situ zircon geochronology

Sample_zircon_spot*	238 U/ 206 Pb age (Ma) (±1 σ)	% ²⁰⁶ Pb* (±1σ)†	Th/U (±1σ)	
SPO20_zircon1	234 (18)	81.8 (2.7)	0.320 (0.003)	
SPO6_zircon2_spot1	187 (23)	58.8 (5.9)	0.161 (0.005)	
SPO10_zircon2	172 (9)	84.4 (1.8)	0.062 (0.002)	
SPO10_zircon7_spot1	163 (8)	95.6 (0.7)	1.079 (0.007)	
SPO10_zircon19a	157 (10)	67.3 (2.6)	0.216 (0.003)	
SPO10_zircon19b	152 (12)	56.7 (2.0)	0.171 (0.003)	
SPO10_zircon1	150 (34)	28.8 (4.8)	0.151 (0.004)	
SPO10_zircon20	148 (9)	87.6 (1.7)	0.962 (0.023)	
SPO6_zircon2_spot2	142 (21)	56.5 (7.3)	0.183 (0.008)	
SPO10_zircon7_spot2	141 (7)	88.1 (1.4)	0.790 (0.004)	
SPO4_z7	140 (3)	99.3 (0.2)	0.136 (0.007)	
SPO20_zircon2	139 (22)	57.9 (5.9)	0.344 (0.010)	
SPO4_z4	131 (4)	99.9 (0.1)	0.120 (0.001)	
SPO10_zirconnz1	126 (10)	68.4 (3.1)	0.288 (0.003)	
SPO4_z1	109 (5)	98.9 (0.8)	0.551 (0.055)	
SPO20_zircon3	97.4 (13)	56.0 (6.5)	0.393 (0.009)	
SPO4_z2	86.1 (2.9)	99.6 (0.2)	0.230 (0.004)	
SPO6_zircon1_spot3	64.3 (0.8)	99.5 (0.1)	1.309 (0.014)	
SPO6_zircon1_spot1	52.5 (1.2)	98.8 (0.1)	0.640 (0.004)	
SPO6_zircon1_spot2	42.4 (0.5)	99.5 (0.1)	1.216 (0.021)	

*Nomenclature is zircon grain number_spot number. (See Figs 14 and 15 for images of the dated grains and concordia diagrams.) †Per cent radiogenic ²⁰⁶Pb.

(Fig. 14). All other results are concordant. The spot on Triassic age zircon is located in a CL-dark interior domain of a euhedral grain, whereas the core of the early Jurassic age zircon appears mottled and has a highly luminescent rim (Fig. 15). The reversely discordant grain (SPO10_zircon 19b; Fig. 14b) is probably due to an overcorrection of common Pb. The youngest age is located in the cracked centre of a zircon that contains regions of bright 1 µm sized inclusions. Spots that overlap these brighter inclusions have higher Th/U (Table 5). The least radiogenic grain yields the results with the largest uncertainty (SPO10 zircon 1, 150.2 ± 34.0 Ma); it should be noted that the uncertainty in these SIMS ages largely depends on the amount of common Pb measured in the grain. Multiple spots were placed on larger individual zircons in samples SPO10 and SPO6. Both spots on zircon grain 7 in sample SPO10 overlap its brighter CL rim, but the spot that overlaps more of the darker CL core is c. 20 myr older. The zircon grain with two spots in sample SPO6 shows homogeneous CL brightness, but spots differ by c. 40 myr in age.

Seven zircons are Jurassic (from 187 ± 23 Ma, SPO6, to $148 \pm$ 9 Ma, SPO10), and the average of all Jurassic results is 161 ± 18 Ma (weighted mean age (WMA) is 160 ± 17 Ma) and is consistent with a single population (mean square weighted deviation (MSWD) of (0.8). It should be noted that we include the WMA as it considers uncertainty in its estimation, and the age that is less precise contributes less to the final age, which is not the case for the average age and uncertainty calculation, which considers all of the uncertainties equally. The most precise Jurassic zircon age is 163 \pm 8 Ma (SPO10) and has a Th/U >1. Seven zircon grains are early Cretaceous, and range in age from 142 ± 21 Ma (SPO6) to $86.1 \pm$ 2.9 Ma (SPO4). The Cretaceous zircon ages average 124 ± 12 Ma with a WMA of 119 ± 3 Ma (MSWD = 13). The spot that yields the oldest early Cretaceous age is located adjacent to the spot that yields a Jurassic age (SPO6). This zircon shows no zoning in BSE or CL. The most precise Early Cretaceous age is 138 ± 6 Ma (sample SPO4), and the average of Early Cretaceous zircon ages is $133 \pm$ 13 Ma (WMA is 134 ± 4 Ma, MSWD 4.4). Four zircons yield ages that average 141 ± 12 Ma (WMA of 140 ± 6 Ma, MSWD 0.01), and these analyses are of either whole grains or rims. Two zircon grains are Late Cretaceous $(97.4 \pm 13.4 \text{ Ma rim analyses}, \text{SPO20},$ and 86.1 ± 2.9 Ma, SPO4).

Discussion

Tectonic environment

The Spongtang ophiolite has a debated formation history (Reuber *et al.* 1983, 1990; Reuber 1986*a*, *b*; Pedersen *et al.* 2001; Mahéo *et al.* 2004). The major element geochemical results we obtained for spinel and olivine in harzburgite sample SPO25 suggest that it is a peridotite with mid-ocean ridge affinity, the product of partial melting of a MORB source beneath a spreading centre with no discernible subduction influence (Fig. 8). This scenario is consistent with the REE patterns of Spongtang clinopyroxene (Fig. 9). With their LREE-depleted patterns, they show no evidence of late-stage melt impregnation nor the extremely great extents of partial melting associated with some SSZ settings (see Pirard *et al.* 2013; Dygert & Liang 2015).

However, an SSZ setting influence on the ophiolite cannot be excluded, on the basis of samples recovered by Mahéo *et al.* (2004), who reported four Spongtang dunites that have spinel Cr#s implying evident SSZ affinity (Cr# >0.6; Fig. 8). We requested the location of these samples to compare with SPO25, but the data are unavailable. We note that Mahéo *et al.* (2004) reported data for lherzolites and harzburgites consistent with our results (Cr# <0.4). The Spongtang ophiolite is commonly assumed to be solely a remnant of an SSZ setting (e.g. Aitchison *et al.* 2000; Baxter *et al.* 2010; Aitchison & Davis 2004; Clift *et al.* 2014), but data reported here suggest a multistage origin.

The seemingly inconsistent Cr#s and REE patterns suggest that the ophiolite contains peridotites from multiple tectonic settings, or that it evolved in a setting that experienced increasing subduction influence through time. The occurrence of peridotites with abyssal and SSZ affinity in one locality has been reported elsewhere (e.g. Choi *et al.* 2008; Dilek *et al.* 2008; Uysal *et al.* 2012; Kingson *et al.* 2017). The record of multiple tectonic settings may be attributed to complex melt–rock reaction (e.g. Elthon 1992; Michibayashi *et al.* 2009; Dygert *et al.* 2016) and/or re-melting and re-crystallization processes that affect mineral compositions and modes as melt ascends (e.g. Niu *et al.* 1997). Such processes may be expected to influence dunites in particular, which are interpreted to represent conduits for mantle melt extraction (e.g. Kelemen *et al.* 1995, 1997;

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Suhr 1999) and have been shown to host multiple generations of melt with distinct source compositions (e.g. Dygert *et al.* 2016).

Cooling history

Geothermometers rely on a temperature-sensitive diffusive exchange of an element or component among phases. Because re-equilibration rates of mineral components and trace elements (e.g. trivalent REE v. divalent major elements) may differ by many orders of magnitude (e.g. Cherniak & Dimanov 2010, and references therein), differences in temperatures given by distinct thermometers for rocks that experienced cooling are expected (see Dygert et al. 2017). For samples that underwent slow, continuous cooling, thermometers based on relatively fast-diffusing major elements (e.g. the olivinespinel of Fabriès 1979, and two-pyroxene solvus of Wells 1977; Brey & Köhler 1990; Putirka 2008) record lower temperatures than those based on slow-diffusing trivalent elements (e.g. the REE-intwo-pyroxene of Liang et al. 2013, and Al-in-orthopyroxene of Witt-Eickschen & Seck 1991). As a general rule, the larger the difference between a temperature given by a trivalent element-based thermometer and olivine-spinel or two-pyroxene solvus thermometer, the slower the sample cooled. Thus, samples that have $T_{\text{REE}} \gg T_{\text{BKN}}$ (which plot far to the left in the $T_{\text{REE}}-T_{\text{BKN}}$ space shown in Fig. 10) cooled slowly relative to samples that have $T_{\text{REE}} \sim T_{\text{BKN}}$ (which plot along or near the blue 1:1 line in Fig. 10).

Sample SPO25 has relatively high and consistent temperatures from trivalent element-based thermometers (990-1058°C) and relatively low (and somewhat scattered) temperatures from solvusbased two-pyroxene thermometers (656-800°C) and olivinespinel thermometers (794–841°C) (Table 4). The inconsistency among the two-pyroxene solvus temperatures may reflect calibration issues, as these thermometers are extrapolated to temperatures well beyond their calibration datasets. The disparity notwithstanding, $T_{\rm BKN}$ and $T_{\rm REE}$ are at the low end of the field defined by ophiolitic peridotites (Fig. 10a). The difference between the T_{REE} and $T_{\rm BKN}$ values is as large as measured in any other ophiolitic peridotite and suggests that Spongtang had an exceptionally long and slow cooling history. Inference of a meaningful cooling rate from the temperature data requires that SPO25 was not disturbed by metasomatic or melt-rock reaction events; these are not suggested by the major and trace compositions of Spongtang pyroxenes (Fig. 9, Tables 2 and 3), or the T_{REE} inversion. Petrographic images of the sample SPO25 suggest alteration (Fig. 6), but our analyses are of intact grains that show largely homogeneous compositions (olivine, clinopyroxene; Table 2) or preservation of zoning (spinel, orthopyroxene; Table 3).

An estimate of the cooling rate of the sample can be made using a modified form of the Dodson (1973) closure temperature equation (Ganguly & Tirone 1999). Assuming an orthopyroxene grain radius of 0.5 mm (consistent with measured grain sizes in SPO25), we calculated cooling curves using methods of Dygert & Liang (2015); the cooling curves are plotted as magenta lines in Figure 10b. We assume that T_{REE} can be modelled using REE diffusion rates in orthopyroxene (Cherniak & Liang 2007; Yao & Liang 2015), and that T_{BKN} can be modelled using rates of Fe–Mg interdiffusion in clinopyroxene (Dimanov & Wiedenbeck 2006; Dygert & Liang 2015). Cooling rates in degrees per year are shown above the highest cooling curve; the position of SPO25 in this temperature space suggests that the Spongtang mantle section cooled at a rate of 100°C Ma⁻¹ or slower.

Such slow cooling is consistent with but at the low end of rates expected for conductive cooling of the mantle lithosphere formed beneath an oceanic spreading centre with a 'normal' 5 km thick crustal section (Dygert *et al.* 2017). The rate may alternatively reflect extended cooling that occurred during or after the closure of a Tethyan basin, and/or the tectonic emplacement of the ophiolite.

The lower $T_{\rm BKN}$ and $T_{\rm REE}$ of SSZ affinity ophiolites (Fig. 10) may partially reflect lower initial temperatures associated with their evolution in relatively cold, hydrous environments. However, we do not view a cooling history as a diagnostic indicator of tectonic affinity because of the possibility of resetting in the events leading up to obduction.

Chronology

The Spongtang ophiolite has been challenging to date because of its multi-stage history coupled with variable alteration. Here we discuss the U–Pb zircon and 40 Ar/ 39 Ar whole-rock ages by grouping results in time and focusing on their possible tectonic history by evaluating information provided by petrography and CL imagery of the dated grains.

A series of Triassic ages is recorded by ⁴⁰Ar/³⁹Ar data from the last degassing steps in basalt subvolcanic sample SPO17 (205 \pm 3 Ma) (Fig. 13) and the darker CL core of a zircon crystal in deformed diabase dyke sample SPO20 of 234 ± 18 Ma (Fig. 15). Sample SPO17 was collected from the upper crust of the Spongtang ophiolite and shows minimal alteration (Fig. 4), whereas sample SPO20 is a highly foliated deformed dyke from the upper mantle unit (Fig. 5). Older Carboniferous to Permian ⁴⁰Ar/³⁹Ar inverse isochron and step ages are found in samples SPO6, SPO10 and SPO22, but these are highly uncertain. Data from last steps are the closest approximation to the high-temperature retention of Ar isotopes in high melting temperature minerals such as pyroxene and plagioclase. It should be noted that previous researchers have also reported Triassic argon ages from amphibole separates from the area, but these are also highly uncertain (Reuber et al. 1989, 1990; Mahéo et al. 2004) (Fig. 3).

Triassic crystallization ages appear inconsistent with the interpretation that the ophiolite formed during the Early Jurassic based on zircon fractions separated from a diorite pod $(177 \pm 1 \text{ Ma};$ Pedersen *et al.* 2001). Triassic ⁴⁰Ar/³⁹Ar whole-rock ages may represent excess ⁴⁰Ar, and the disturbed age spectra suggest this as a possibility. Also, the Triassic zircon is discordant (Fig. 14c), and the older core may have been entrained in the rock from a different source, potentially continental crustal fragments that were adjacent to the northern passive continental margin of Greater India (Cao *et al.* 2018). However, the tectonic setting of the Early Jurassic age diorite pod (Pedersen *et al.* 2001) suggests that the sample intrudes higher-level gabbros of the ophiolite, and thus it may cut older rocks.

We report four zircon ages that overlap with the reported Jurassic age, including diabase dyke sample SPO6 $(187 \pm 23 \text{ Ma})$ and plagiogranite SPO10 $(172 \pm 9, 163 \pm 8 \text{ and } 150 \pm 34 \text{ Ma})$ (Table 5, Fig. 15). It should be noted that sample SPO6 has a previously reported K/Ar age of 169.0 ± 9.5 Ma (Reuber *et al.* 1989, 1990). The average of all Jurassic zircon ages is consistent with a single age population and average 161 ± 18 Ma (WMA of 160 ± 17 Ma, MSWD of 0.8). The most precise Jurassic zircon age is located on a SIMS spot that overlaps a darker core and brighter rim (163 ± 8 Ma, SPO10; Fig. 15). The zircon with the Jurassic age in SPO6 shows flat CL zoning, but an adjacent spot is *c.* 40 myr younger (142 ± 21 Ma) and is similar to the age of the rim of the other Jurassic-age zircon in SPO10 (141 ± 7 Ma).

Overall, the Triassic to Late Jurassic ages suggest that the Spongtang ophiolite probably began its origin as part of the oldest portion of the Neotethyan domain (Late Triassic to Late Cretaceous; Sinha-Roy 1982; Şengör & Atayman 2009; Cao *et al.* 2018). This region is sometimes termed the Ceno-Tethyan Ocean (Metcalfe 1999; Matsuoka *et al.* 2002; Wakita & Metcalfe 2005). The Yarlung Zangbo Suture Zone itself is mapped as the zone of the closure of the Ceno-Tethyan Ocean (e.g. Metcalfe 1999, 2009, 2013). Based on the thermometry and geochemical data, we model the origin of

the Spongtang ophiolite as part of a spreading centre with a spreading rate $>2 \text{ cm a}^{-1}$, as spreading centres with slower spreading rates have thin or no crust (e.g. Cann *et al.* 1997; Dick *et al.* 2003, 2008; Morris *et al.* 2009; Craig & Parnell-Turner 2017). The ophiolite cooled rapidly by hydrothermal circulation and tectonic exhumation, recording higher T_{REE} compared with T_{BKN} (Fig. 10; Dygert & Liang 2015).

Both whole-rock ⁴⁰Ar/³⁹Ar and zircon ages are reported from samples SPO4, SPO6 and SPO10. The expectation is that zircon ages should be older than those obtained using the ⁴⁰Ar/³⁹Ar approach, with the last degassing argon steps approaching zircon U-Pb results. The larger uncertainty in the zircon ages in samples SPO6 and SPO10 preclude this type of interpretation, although the oldest zircon in sample SPO4 (140 ± 3 Ma) overlaps the last degassing ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ step-heating results (142.8 ± 16.6 Ma). Reuber et al. (1989, 1990) also reported a 139 ± 8 Ma K/Ar age from amphibole in this sample, suggesting that the c. 140 Ma result should be considered part of the history of the ophiolite. Overall, we report 10 Cretaceous zircon ages from the Spongtang ophiolite, and four of these zircons yield ages that average 141 ± 12 Ma (WMA of 140 ± 6 Ma, MSWD <0.01). Spots on two of these zircons overlap rims in CL, and the textural relationships of the Cretaceous zircons suggest that they could be associated with a tectonic event that facilitated zircon recrystallization after the Jurassic.

The period recorded by the Cretaceous zircons has been ascribed to the development of an intra-oceanic arc represented by the Samail (Oman), Zagros (Iran), Chagai-Raskoh (western Pakistan, southern Iran, eastern Afghanistan), Kandahar and Kohistan-Ladakh, Dazhuqu and Zhongba arcs (Brookfield & Reynolds 1981; Bhutani et al. 2004; Abrajevitch et al. 2005; Dai et al. 2011; Siddiqui et al. 2012; Baxter et al. 2016) (Fig. 1). Reuber et al. (1989, 1990) suggested that a significant intra-oceanic thrusting event affected the Spongtang ophiolite during the Early Cretaceous, and Mahéo et al. (2004) suggested that this was a time when subduction was initiated, based on interpretations of disturbed ⁴⁰Ar/³⁹Ar amphibole step-heating data. The Early Cretaceous (140–130 Ma) is also a time when the India subcontinent initiated rifting from other continents and fragments associated with Gondwana (e.g. Scotese et al. 1988; Jadoul et al. 1998; Hu et al. 2010).

A Late Cretaceous intra-oceanic arc has been suggested to be present within the Neotethyan Ocean near the palaeo-equator (e.g. Reuber 1986b; Abrajevitch *et al.* 2005; Metcalfe 2009; Dai *et al.* 2011; Siddiqui *et al.* 2012, 2017) or *c.* 30°N (Zhu *et al.* 2013). Paleogeographic reconstructions of Lawver *et al.* (2018) restrict its location to the south of the Lhasa Terrane. It should be noted that some have indicated that the Lhasa Terrane had already accreted onto other Tibetan-related continental fragments by the Late Jurassic–Early Cretaceous (Rolland 2002; Kapp *et al.* 2003, 2007; Guynn *et al.* 2006; Zhu *et al.* 2013), and others have suggested the presence of multiple arcs (e.g. Zyabrev *et al.* 2008). This observation and alternative palaeogeographical reconstructions (e.g. Ali & Aitchison 2005) do not change our overall observation that the data reported here are consistent with the presence of a Late Cretaceous intra-oceanic Neotethyan arc.

We speculate that the Spongtang ophiolite records a transition from mid-ocean ridge affinity-type spreading centre to a setting with increasing subduction influence, perhaps after a subduction polarity reversal as the Neotethyan basin closed before obduction. This scenario is consistent with the presence of both the abyssal and the SSZ signatures, with the Spongtang ophiolite exposures and the seafloor spreading suggested by the REE and solvus thermometry, and trace element geochemistry. The timing of this transition cannot be specifically constrained with the ages reported here, but we speculate that this occurred sometime after the Jurassic, and is probably recorded by ages of zircon rims (*c.* 140 Ma).

Late Cretaceous zircon ages may represent continued crystallization as the subduction-zone influence increased, or are the result of the mixing of age domains. It should be noted that these results are similar to the ⁴⁰Ar/³⁹Ar age of basalt associated with the Indus Suture north of the field area $(128.2 \pm 2.6 \text{ Ma; Bhutani et al. 2004})$ and therefore are likely to represent continued crystallization and emplacement. Zircon ages from SPO4 (131.2 ± 4.3 Ma) and SPO10 $(126.3 \pm 10.2 \text{ Ma})$ overlap zircon ages from underlying sediments $(129 \pm 7 \text{ Ma}; \text{Clift et al. 2014})$ (Table 5). The spot that yields the SPO4 age samples multiple zircon clusters and the spot on the zircon in SPO10 overlaps a darker CL rim with veins that intrude its bright core (Fig. 15). These ages thus probably represent mixed age domains. Inverse isochron 40Ar/39Ar Late Cretaceous ages are highly uncertain, but selected step heating ages are more precise $(101.7 \pm 1.2 \text{ Ma}, \text{ dyke SPO15}; 98.4 \pm 2.3 \text{ Ma}, \text{ basalt subvolcanic}$ sample SPO17; 73.7 ± 2.2 Ma, pillow lava SPO19) (Figs 12 and 13). These ages overlap the oldest reported age constraints timing the collision of India and the Kohistan-Ladakh island arc during the Late Cretaceous (c. 95-65 Ma) (see review by Khan et al. 2009). The Late Cretaceous has been considered to the time of ophiolite obduction in the region $(82 \pm 6 \text{ Ma}; \text{Brookfield & Reynolds 1981})$ and the crystallization of other adjacent ophiolite sequences that may be correlative $(80.2 \pm 1.5 \text{ Ma}, \text{ Muslim Bagh}; \text{ Kakar et al.}$ 2012). Some of the Late Cretaceous ages we report are similar to the oldest ages reported for the Kohistan and Kargil batholiths, part of a granitic system north of the field area that has been used to time Neotethyan subduction (c. 101 Ma; Honegger et al. 1982; Schärer et al. 1984). However, the majority of zircon U-Pb ages reported for those granites are younger (see Weinberg & Dunlap 2000; Khan et al. 2009).

The oldest Cenozoic ⁴⁰Ar/³⁹Ar ages we obtained are from the last degassing steps of samples SPO14 (58.7 ± 0.7 Ma) and SPO10 $(56.7 \pm 13.6 \text{ Ma})$ (Fig. 12), and the inverse isochron age of the last steps of sample SPO19 (54.76 ± 0.54 Ma) (Fig. 13). These $^{40}\mathrm{Ar}/^{39}\mathrm{Ar}$ ages average 56.7 \pm 5.2 Ma (WMA of 56.2 \pm 0.2 Ma, MSWD 6.6). The oldest Cenozoic zircon age is from a single zircon in sample SPO6 (64.3 \pm 0.8 Ma), and the grain has younger age domains of 52.5 ± 1.2 and 42.4 ± 0.5 Ma (Fig. 15, Table 5). BSE images of the dated zircon with bright and cracked zones suggest that the results probably time metamorphism and zircon recrystallization. The ages are consistent with palaeomagnetic data indicating a decrease in speed of the Indian craton at this time (e.g. Klootwijk et al. 1992; Molnar & Stock 2009; Copley et al. 2010; Dupont-Nivet et al. 2010; Cande & Stegman 2011; van Hinsbergen et al. 2011), and the proposed timing of Spongtang ophiolite obduction (70-60 Ma, Searle 1986; 55-50 Ma, Kelemen et al. 1988; Garzanti et al. 2005; Baxter et al. 2016). All Cenozoic ages we obtain are in line with broad estimates of the timing of Indo-Asia collision (e.g. Rowley 1996; Yin & Harrison 2000; Guillot et al. 2003; Aitchison & Davis 2004; Khan et al. 2009; Bhutani et al. 2009; Molnar & Stock 2009; Cai et al. 2011; van Hinsbergen et al. 2011; Hu et al. 2016; Najman et al. 2017; Tong et al. 2017).

Although uncertain, five of the Spongtang 40 Ar/ 39 Ar determinations produced Oligocene absolute inverse isochron ages (c. 34 Ma, SPO4, SPO6, SPO22, SPO10 and SPO14; Figs 11 and 12). Others have suggested that a dual-collision scenario occurred in the Himalayas during the Oligocene (e.g. Aitchison *et al.* 2000, 2007; Aitchison & Davis 2004; Abrajevitch *et al.* 2005; Ali & Aitchison 2005, 2014). One sample yields a more precise Miocene result (21.2 ± 2.4 Ma; Fig. 13). It should be noted that Oligocene to Miocene ages are commonly reported throughout the Himalayas and may time events related to mountain-building along large-scale thrust systems in the range (see review by Yin 2006). These Oligocene ages are uncertain and could represent the mixing of age domains coupled with excess argon. We choose to discuss the results as a significant number of dated samples seem to have this

age signature. The youngest ⁴⁰Ar/³⁹Ar results (2–5 Ma) recorded by all samples probably represent the degassing of alteration minerals (muscovite, sericite, chlorite).

Conclusions

Here we report geochemical and geochronological data from samples from the Spongtang ophiolite, a piece of obducted oceanic lithosphere located in the NW Himalayas. Major and trace element mineral chemistry and geothermometry suggest that the ophiolite experienced slow cooling after forming at a spreading centre with mid-ocean ridge affinity. The difference between the T_{REE} and T_{BKN} values estimated for harzburgite sample SPO25 is as large as that for any ophiolitic peridotite yet analysed, and suggests an exceptionally long and slow cooling history, assuming the lack of disturbance by metasomatic or melt-rock reaction events. The geochemical and geochronological results reported here, coupled with previous work (Pedersen et al. 2001; Mahéo et al. 2004), suggest that the ophiolite began initial crystallization at a spreading centre with a spreading rate >2 cm a⁻¹ within the oldest parts of the Neotethyan ocean possibly during the latest Triassic to early Jurassic.

The Triassic time frame is constrained by the highest temperature $^{40}\mathrm{Ar}/^{39}\mathrm{Ar}$ degassing step of a basalt subvolcanic assemblage (205 \pm 3 Ma) and a single ²³⁸U-²⁰⁶Pb age of a zircon core in a deformed upper mantle unit dyke (234 ± 18 Ma). The Jurassic timing is the result of multiple 238 U $-^{206}$ Pb zircon ages (161 ± 18 Ma, WMA of 160 ± 17 Ma) that are c. 10 myr younger, but still similar to those reported previously (177 ± 1 Ma; Pedersen *et al.* 2001). Based on the presence of rocks with the chemistry of rocks associated with SSZ settings (Mahéo et al. 2004), we suggest that the ophiolite then experienced increasing subduction influence during continuing deformation that occurred during the Early Cretaceous. The Early Cretaceous timing is constrained by ²³⁸U-²⁰⁶Pb zircon data (123.6 \pm 12.1 Ma, WMA age of 119.1 \pm 2.5 Ma). We model this transition as a subduction polarity reversal along a subduction zone adjacent to the spreading centre that formed the ophiolite owing to closure of the Neotethyan Ocean. This process may have occurred as part of the development of a larger intra-oceanic arc system that has been documented elsewhere across the Neotethyan Ocean. Based on the most precise zircon and whole-rock 40Ar/39Ar ages from several samples, final obduction of the Spongtang ophiolite occurred in the Cenozoic. The ²³⁸U-²⁰⁶Pb zircon ages constrain the timing of obduction between 64.3 ± 0.8 and 42.4 ± 0.5 Ma, which is broadly consistent with whole-rock ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ ages of 56.7 ± 5.2 Ma.

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