

ORIGINAL ARTICLE

Pressure–temperature–time evolution of subducted crust revealed by complex garnet zoning (Theodul Glacier Unit, Switzerland)

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Abstract

Collisional orogens commonly include mono-metamorphic and poly-metamorphic units, and their different evolution can be difficult to recognize and reconcile. The Theodul Glacier Unit (TGU) in the Western Alps consists of an association of meta-sedimentary and metamafic rocks embedded within the Zermatt-Saas tectonic unit. In spite of recent petrological studies, it remains unclear whether these rocks underwent one or multiple metamorphic cycles. In this study, different lithologies from the TGU unit (mafic schist, mafic granofels, and chloritoid schist) were investigated for petrography, quantitative compositional mapping of garnet, thermodynamic modelling, and Lu–Hf garnet dating. The data reveal a coherent mono-metamorphic history with a β -shaped pressure–temperature (P – T) path characteristic of oceanic subduction. Garnet Lu–Hf ages yield a restricted garnet crystallization time window between 50.3 and 48.8 Ma ($\pm 0.5\%$, 2SD). A prograde metamorphic stage recorded in garnet cores yields conditions of $490 \pm 15^\circ\text{C}$ and 1.75 ± 0.05 GPa. Maximum pressure conditions of 2.65 ± 0.10 GPa and $580 \pm 15^\circ\text{C}$ were reached at 50.3 ± 0.3 Ma. Initial exhumation was rapid and led to isothermal decompression to 1.50 ± 0.10 GPa within 1 Ma. This decompression was associated with lawsonite breakdown in mafic schist and in mafic granofels, causing intense fluid–rock interaction within and between different lithologies. This process is recorded in garnet textures and trace element patterns, and in the major element composition of K-white mica. Initial exhumation was followed by re-heating of $\sim 30^\circ\text{C}$ at a pressure of 1.50 ± 0.10 GPa. Perturbation of the subduction-zone thermal structure may be related to upwelling of hot asthenospheric mantle material and transient storage of the unit at the crust–mantle boundary.

KEYWORDS

Lu–Hf garnet whole-rock geochronology, phase equilibria, quantitative compositional mapping, Zermatt-Saas Unit, Zr-in-rutile thermometry

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1 | INTRODUCTION

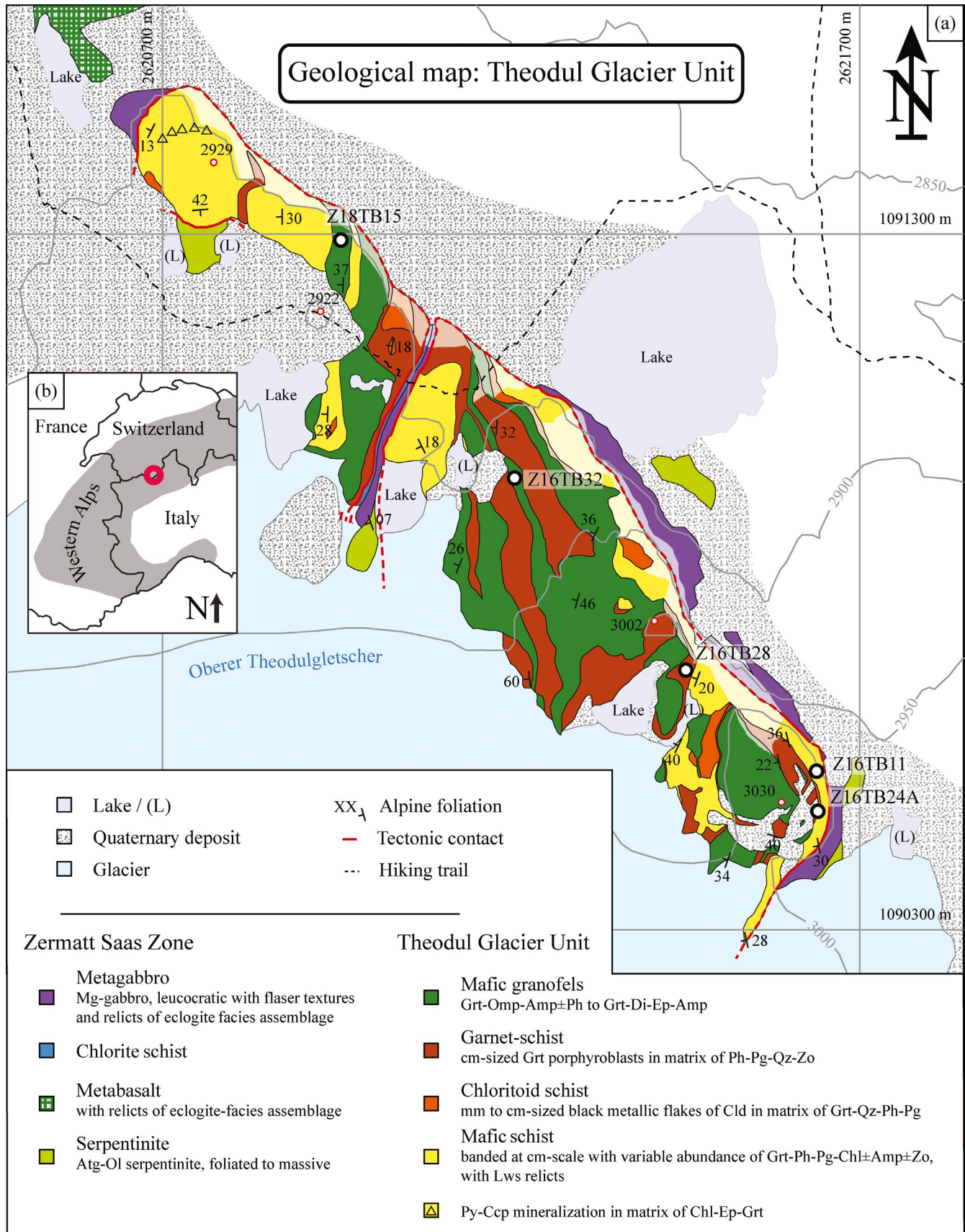
In poly-metamorphism terranes, several metamorphic cycles are experienced by a rock and each metamorphic cycle involves a sequence of prograde and retrograde transformations affecting the rock texture and mineral compositions. However, the transformation is rarely complete and mineral relics are key to decipher multi-stage events as well as pressure and temperature (P – T) evolution. Garnet is a common rock-forming mineral known to be resilient to re-equilibration (Baxter et al., 2013) and therefore is a good recorder of poly-metamorphism (Angiboust et al., 2017; Feenstra et al., 2007; Giuntoli et al., 2018; Lanari et al., 2017; Putnis, 2009; Thiessen et al., 2019). The study of garnet texture and chemical composition gives insights into its growth history. When temperatures are low enough to prevent intracrystalline diffusion, continuous garnet growth over a P – T segment produces a typical chemical zoning witnessing local equilibrium between the garnet crystal and the matrix of the rock at the time of garnet growth (Spear, 1988). For example, Mn strongly partitions into garnet and spessartine is consequently enriched in the core relative to the rim composition and exhibits a bell-shape zoning pattern (Hollister, 1966). On the other hand, a sharp compositional contact between distinct garnet zones within a single crystal is commonly interpreted as reflecting two distinct metamorphic stages or metamorphic cycles (Compagnoni, 1977; Giuntoli, Lanari, Burn, et al., 2018; Le Bayon & Ballèvre, 2004; Manzotti & Ballèvre, 2013), or as a consequence of the infiltration of externally derived fluids (e.g. Angiboust et al., 2014; Jamtveit & Hervig, 1994). Garnet major element composition is additionally influenced by the reactive bulk composition, which is the bulk composition of the reactive volume (e.g. Lanari & Engi, 2017). The reactive bulk composition changes during garnet growth by fractionation (Evans, 2004; Konrad-Schmolke et al., 2011) affecting the isopleth positions in equilibrium phase diagrams (Lanari & Engi, 2017; Spear, 1988). In addition, the same garnet composition can be predicted to be stable under different geotherms (e.g. at High- T –Low- P and Low- T –High- P) for a given bulk composition (see figure 8 in Lanari & Engi, 2017). This will inevitably complicate the interpretation of garnet zoning and the distinction between mono- and poly-metamorphic growth.

Besides its chemistry, garnet geochronology enables time constraints, providing a further key tool to distinguish between mono- and poly-metamorphic growth. Both Lu–Hf and Sm–Nd systematics can be employed as chronometers in garnet, because for both systems garnet solely fractionates parent over daughter forming an unusually high parent/daughter isotope ratio (Baxter et al., 2017). The main advantage of Lu–Hf geochronology is that the diffusivity of Hf is low with respect to Nd, making Lu–Hf systematics less susceptible to be affected by diffusive isotope re-equilibration

(Bloch et al., 2015; Kohn, 2009; Scherer et al., 2000; Smit et al., 2013). During garnet growth, Lu is generally enriched in the garnet core due to the high compatibility of Lu in garnet (e.g. Konrad-Schmolke, O'Brien, et al., 2008). As a consequence, Lu–Hf ages of bulk garnet crystals are typically biased towards the initial stage of garnet growth (Konrad-Schmolke, O'Brien, et al., 2008; Lapen et al., 2003; Otamendi et al., 2002; Skora et al., 2006). However, complex Lu zoning in garnet is sometimes observed, where distinct garnet zones are enriched in Lu. Such element distribution might increase the complexity of age interpretation, especially if garnet growth duration between Lu maxima exceeds analytical precision (Lapen et al., 2003; Skora et al., 2006; Smit et al., 2010).

In the Western Alps, the recognition of, and distinction between, mono- and polycyclic metamorphic terranes has become of crucial importance in tectonic and palaeogeographic reconstructions. Mono-metamorphic units are clearly present in the form of terranes comprising Mesozoic sediments and Jurassic–Cretaceous oceanic crust metamorphosed during the Alpine orogeny (e.g. Zermatt-Saas Zone, ZSZ). Poly-metamorphic units are found or suspected in various locations. These are interpreted as portions of continental basement which, in addition to their Alpine overprint, record Permian, and/or older metamorphism (e.g. the Sesia Zone and Dent Blanche; Manzotti et al., 2014). An example of a unit that is potentially poly-metamorphic, yet not fully identified as such, is the Theodul Glacier Unit (TGU)—a metasedimentary sliver embedded within the ZSZ (Figure 1) that has been described as a coherent unit (Weber & Bucher, 2015). This unit largely preserves an Alpine, high- P assemblage, but was recently interpreted as poly-metamorphic, based on the investigation of garnet textures and thermodynamic modelling (Bucher et al., 2019, 2020). The TGU has been therefore interpreted as a small continental outlier embedded in the oceanic units of the ZSZ. Alternatively, the unit could be mono-metamorphic, representing a sedimentary sequence deposited on the oceanic crust of the ZSZ ophiolite, much alike the metasediments of the Lago di Cignana locality (Reinecke, 1998).

To better constrain the metamorphic history and tectonic significance of the TGU, various rocks from this unit were subjected to a multi-method analysis. Two potentially poly-metamorphic metasedimentary lithologies were studied and their results compared with those from metamafic rocks, with the latter being used as benchmark to characterize the mono-metamorphic character of the P – T path of this unit. The mono- versus poly-metamorphic character of garnet in the metasedimentary rocks was assessed based on petrological observations, quantitative compositional mapping, thermodynamic modelling, and Lu–Hf chronology. Additional compositional mapping and thermodynamic modelling was performed on phengite. Zr-in-rutile content and chloritoid



stability were used to constrain temperature conditions at specific metamorphic stage. Our results show that the TGU is a mono-metamorphic unit that followed a β -shaped P – T path during the Eocene Alpine metamorphic cycle.

2 | GEOLOGICAL AND TECTONIC SETTINGS

The Penninic domain in the Western Alps is a complexly folded nappe stack of rocks with continental and oceanic origin that have different metamorphic histories and were assembled during the Alpine orogeny. The ZSZ is a south Penninic tectonic unit made of Mesozoic ophiolites that underwent UHP and eclogitic HP metamorphism (Bearth, 1967; Dal Piaz & Ernst, 1978; Reinecke, 1991) during Eocene (Amato et al., 1999; De Meyer et al., 2014; Duchêne et al., 1997; Lapen et al., 2003; Rubatto et al., 1998). It has been interpreted as being an exhumed relict of the Piemont–Ligurian oceanic lithosphere with mid-ocean ridge affinity (Bearth, 1967), which is now sandwiched between the underlying Briançonnais basement (e.g. the Monte Rosa and the Gran Paradiso massifs) and the overlying Austroalpine units (e.g. the Sesia–Dent Blanche nappe system; Escher et al., 1997). The ZSZ is made of an association of ultramafic rocks, mafic rocks, and metasediments, representing a dismembered ophiolitic sequence (Bearth, 1967). Whether the ZSZ is a coherent piece of oceanic lithosphere, which was then only partially disrupted during exhumation (Angiboust et al., 2009), or whether it is a tectonic mélange of different tectonic slices (Bousquet et al., 2008; Groppo et al., 2009; Negro et al., 2013) is still unclear.

The ophiolites of the ZSZ consist of metamorphosed metabasalts and metagabbros of Jurassic protolith age (Beltrando et al., 2010; Pleuger et al., 2007; Rubatto et al., 1998; Rubatto & Hermann, 2003) situated on top of a thick section of serpentinite (Fassmer et al., 2016; Kirst & Leiss, 2016; Pleuger et al., 2007). The oceanic sedimentary cover—also referred to as *Schistes Lustrés* or *Bündnerschiefer*—is mainly composed of marbles, schists, calcareous schists, meta-radiolarites, and Mn-bearing meta-cherts with a Jurassic to Cretaceous protolith age (Bearth, 1967). The sediments that overlie the metabasalts and metagabbros form a thin continuous cover of the oceanic crust (Pal Piaz, 1979). The overall stratigraphy shows slow spreading ridge affinity (Dilek & Furnes, 2014).

Three kilometre-thick slices of continental crust that were reworked at high- P conditions occur discontinuously throughout the ZSZ section, especially in its uppermost part: Etirol-Levaz, Glacier Rafray, and Monte Emilius. They are either localized within the ophiolitic nappe of the ZSZ (Angiboust et al., 2017; Ballèvre et al., 1986; Beltrando et al., 2010; Dal Piaz, 2001; Fassmer et al., 2016; Kirst & Leiss, 2016) or along the tectonic contact between the Combin zone and ZSZ

(Dal Piaz et al., 2001). Two different origins have been proposed for these continental slices: (a) They are slivers of one of the adjacent continental units, which were incorporated into the ZSZ during the Alpine orogeny (Kirst & Leiss, 2016; Pleuger et al., 2007). This interpretation explains the similarities with the typical characteristics of the Sesia-Zone and the Dent Blanche nappe (Pleuger et al., 2007). (b) They are fragments of continental crust that were dismembered during Permian to Jurassic extension to form extensional allochthons over the Mesozoic oceanic crust (Beltrando et al., 2010; Dal Piaz, 2001). In the latter case they would form a palaeogeographic entity with the ZSZ zone (Fassmer et al., 2016). A similar close relationship between continental basement slices and serpentinites is commonly found along ocean–continent transition zones, where movements along detachment faults dipping underneath the thinned continental margin result in the tectonic erosion of small pieces of the margin (Beltrando et al., 2010, 2014; Péron-Pinvidic & Manatschal, 2009; Vitale Brovarone et al., 2011).

The TGU is located SW of the village of Zermatt in Switzerland, close to the Trockener Steg cable car station. The outcropping rocks show little weathering as the retreat of the ‘Oberer Theodulgletscher’ occurred only during the last three decades. The TGU is an association of metasedimentary and metamafic rocks embedded within the ZSZ unit (Figure 1), which has been mapped as a coherent unit (Bovay, 2021; Weber & Bucher, 2015). The mafic rocks from TGU show within-Plate Basalt chemistry (Weber & Bucher, 2015). Thermodynamic modelling of the eclogitic rocks based on garnet chemistry suggests a mono-metamorphic history with a peak metamorphism at 2.20 ± 0.10 GPa and $580 \pm 50^\circ\text{C}$, slightly lower than those of the ZSZ eclogites (Weber & Bucher, 2015). Garnet Lu–Hf dating of mafic rocks records prograde metamorphic mineral growth and yields an Alpine age at 58–56 Ma (Weber et al., 2015), which is slightly older than the peak metamorphism of ZSZ dated between 50 and 40 Ma (De Meyer et al., 2014; Dragovic et al., 2020; Duchêne et al., 1997; Lapen et al., 2003; Rubatto et al., 1998). On the other hand, zircon U–Pb dating yields an age of 295 ± 16 Ma, which was interpreted to date pre-Alpine granulite facies metamorphism (Bucher et al., 2020). The metasedimentary rocks have been described as an association of garnet–phengite schists and garnet–biotite schists (Weber & Bucher, 2015). In this work, detailed field investigation (Figure 1) resulted in the subdivision of the garnet–phengite schist lithology into two lithological units: garnet schist and mafic schist. In addition, the garnet–biotite schists are renamed as chloritoid schist (Cld-schist), consistent with the observed mineral assemblage (see below). In summary, the highly foliated metasedimentary rocks consist of an intricate imbrication of mafic schist, garnet schist, and Cld-schist that constitute a coherent unit. Large garnet porphyroblasts in the garnet schist recorded two metamorphic stages: a pre-Alpine

granulitic core stable at ~ 0.7 GPa and $\sim 780^\circ\text{C}$ is overgrown by an Alpine rim generation formed at ~ 1.7 GPa and $\sim 530^\circ\text{C}$ (Bucher et al., 2019). Cation diffusion across the contact between the two garnet generations occurred at $530\text{--}580^\circ\text{C}$ and relaxed zoning for *c.* 10 Ma (Bucher et al., 2019).

3 | PETROLOGICAL DESCRIPTION

To better constrain the *P–T* conditions experienced by the TGU, we performed a petrological study focusing on mafic granofels and metasedimentary schists. In this study, three types of lithologies are investigated: the mafic granofels, the mafic schist, and the Cld-schist.

3.1 | Mafic granofels

Mafic granofels occur either as large bodies of 10 m wide and 10–100 m long, or as smaller, stretched boudins of various sizes (cm to m) embedded within the surrounding schists (Figure 2a). This variable lithology can be described by two characteristic types of mineral assemblage (Table 1) (Figure 2b,c). The first type of TGU mafic granofels (sample Z18TB15) has an assemblage of garnet–omphacite–amphibole±phengite (Figure 2b,d,e), similar to that described by Weber and Bucher (2015). The second rock type (sample Z16TB32) shows an assemblage of garnet, diopside, zoisite and/or zoisite–epidote solid solution, and amphibole (Figure 2c). Due to the highly variable mineral modal composition of the mafic rocks, in the field they range in colour from milky to dark green (Figure 2a–d). Garnet grains are mostly euhedral with diameters between 1 and 5 mm and may show textural equilibrium with the amphiboles (Figure 2e). Mafic boudins exhibit undeformed cross-cutting veinlets of garnet–diopside–amphibole–quartz–calcite and show a continuous layer of euhedral garnet at the contact with the host rock on both vein sides. Retrogression locally replaced the main HP mineral assemblage with chlorite, amphibole, epidote, paragonite, albite, and titanite. According to Weber and Bucher (2015), the chemical composition of the different population of amphiboles are either unzoned glaucophane porphyroblasts or Ca-amphibole with actinolitic core and barrositic–edenitic rim. Retrogressed domains in mafic granofels show abundant albite–epidote veins.

3.2 | Mafic schist

A variety of mafic schists are found interlayered at the centimetre to metre scale. The layers are oriented parallel to the main foliation and the transition from one type to another can be either gradual (mm to cm) or sharp (Figure 2f). The

mineral assemblage common to every type of mafic schist is garnet–phengite–paragonite–quartz–chlorite (Figure 2g), with additional zoisite and amphibole (Table 1). Mineral abundance can vary considerably; for example, sample Z16TB11 comprises <1 vol.% of zoisite and 15% amphibole, whereas sample Z16TB28 contains 10%–35% zoisite and up to 5% amphibole. Pseudomorphs after lawsonite were observed in every type of mafic schist (Figure 2f). These occur as whitish diamond-shaped mineral aggregates of up to 1 cm and are composed of zoisite–quartz–paragonite, sometimes with small euhedral garnet. Three main garnet populations are observed in the mafic schists (Figure 2g). (a) Sub-euhedral grains in the mineral matrix, which are up to 200 μm in diameter and are mostly inclusion free. These occur either as banding of millimetre to centimetre scale in which each grain is stretched along the main foliation, or as clusters. This garnet type preserves complex zoning with relict cores and a distinct chemical composition from the other garnet types (see below). (b) Euhedral grains of 0.2–0.7 mm in diameter; these matrix garnet do not contain relict cores. (c) Euhedral garnet of 1–7 mm in diameter, which are rich in inclusion of various minerals (zoisite, rutile, titanite, apatite, pyrite, amphibole, quartz, zircon, and white mica). Inclusion textures in garnet vary from sample to sample and indicate syn to post kinematic entrapment. Greenschist facies retrogression caused local growth of chlorite and amphibole rims around garnet, titanite rims around rutile, and production of minor poikilitic albite.

3.3 | Chloritoid schist

This lithology was previously mapped by Weber and Bucher (2015) as garnet–biotite schists. Both in the field and in thin section, biotite was not identified. Instead, abundant chloritoid flakes were observed along the Alpine foliation, hence here the term Cld-schist is adopted. This lithology (sample Z16TB24A) is typified by black metallic flakes of chloritoid in a foliated matrix of garnet, quartz, phengite, and paragonite (Figure 2h,i, Table 1). Rare altered amphibole porphyroblasts and elongated rutile are also present. Garnet grains are euhedral with a diameter of 0.2–0.4 mm and occasionally show an atoll-shaped texture. The lithology contains garnet–chloritoid–quartz veins.

4 | ANALYTICAL METHODS

4.1 | Scanning electron microscope and electron micro probe

Sample imaging (backscattered electron mode) was performed using a scanning electron microscope (SEM) ZEISS

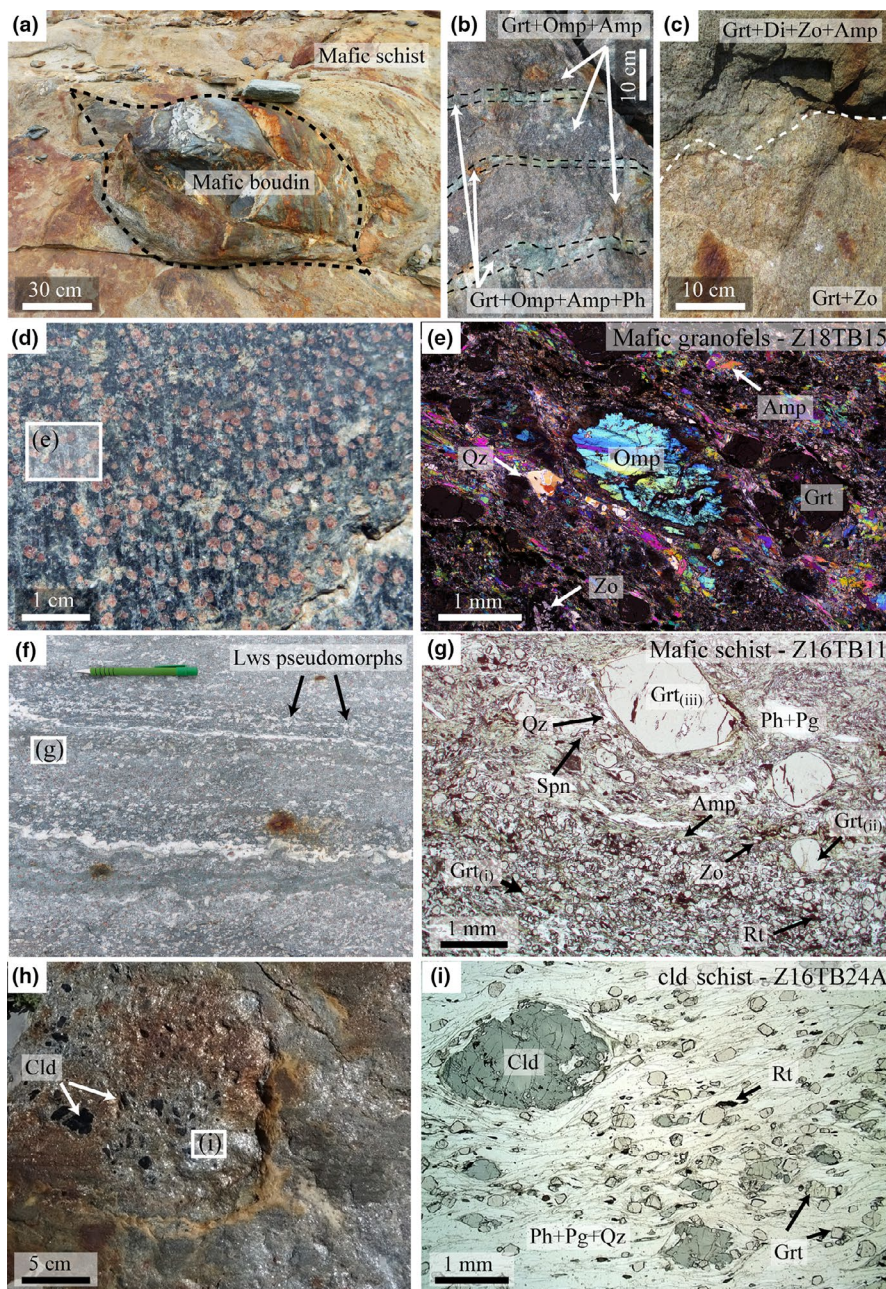


FIGURE 2 Outcrop photographs and sample microphotographs from the representative lithologies of the TGU. Mineral abbreviations follow Whitney and Evans (2010). White squares in (d), (f), and (h) are representative location of indicated microphotograph. (a) Mafic boudin stretched within the schists. (b) Outcrop view of banded mafic granofels. (c) Outcrop view of mafic granofels with smooth transition between Di-rich to Zo-rich domain. (d) Outcrop view of mafic granofels: lighter areas are predominantly Omp-rich, whereas darker areas are Amp-rich. (e) Microphotograph of mafic granofels Z18TB15 showing textural equilibrium between Omp and Grt, both surrounded by Amp. (f) Outcrop view of mafic schist. Layering is parallel with the well-developed foliation. White diamonds-shaped domains are interpreted as pseudomorphs after lawsonite. (g) Scanned thin section of mafic schist Z16TB11 showing transition from a zone rich in small garnet (<200 μm , bottom of image) to a zone richer in garnet porphyroblasts. (h) Outcrop view of a Cld-schist, with dark flaky chloritoid. (i) Scanned thin section image of Cld-schist Z16TB24A: the foliation is marked by Ph+Pg+Qz

EVO50 at the Institute of Geological Sciences (University of Bern) with 20 keV acceleration voltage and 1 nA beam current.

Quantitative chemical analyses of garnet, white mica, and chloritoid were performed with the electron probe micro-analyser (EPMA) JEOL JXA-8200 superprobe at the Institute of Geological Science (University of Bern). Spot analyses were performed using 15 keV accelerating voltage, 20 nA specimen current, and 40 s dwell time (10 s for each background after 20 s on peak). Nine oxide components were obtained using synthetic and natural standards: almandine (SiO_2 , Al_2O_3 , and FeO), albite (Na_2O), anorthite (CaO), orthoclase (K_2O), forsterite (MgO), ilmenite (TiO_2), and tephroite (MnO).

Quantitative compositional maps were generated from X-ray intensity maps using spot analyses acquired in the same area as internal standards. The X-ray maps were measured by WDS with 15 keV accelerating voltage and 100 nA specimen current, with various dwell time and resolutions. Nine elements (Si, Ti, Al, Fe, Mn, Mg, Na, Ca, and K) were measured at the specific wavelength in two successive passes (with Na, Ca, and K analysed on the first scan). Compositional maps were processed using XMapTools 3.2.1 (Lanari et al., 2014, 2019). Structural formulas for garnet and phengite were calculated for each map using the external functions available in XMapTools and considering a minimum of 17 internal standards. Representative compositions of each garnet and phengite growth zone were obtained by averaging pixels

TABLE 1 Sample list including geolocalization and mineral assemblages

Sample	Latitude	Longitude	Major phases	Minor phases
Mafic granofels				
Z16TB32	45°58'9.66"N	7°42'42.84"E	Grt+Ep+Di+Amp	Qz+Ttn+Ap
Z18TB15	45°58'20.9"N	7°42'34.8"E	Omp+Grt+Amp	Ep+Qz+Ph+Pg+Chl+Ap+Rt+Ttn
Mafic schist				
Z16TB11	45°58'1.04"N	7°43'0.53"E	Grt+Qz+Ph+Pg+Amp+Ep	Chl+Ab+Rt+Ttn+Ap
Z16TB28	45°58'1.26"N	7°42'56.64"E	Grt+Ep+Chl+Qz+Ph	Ttn+Rt+Ab+Ap
Chloritoid schist				
Z16TB24A	45°57'54.72"N	7°43'4.62"E	Grt+Qz+Ph+Pg+Cld+Amp	Chl+Rt+Ttn

from manually selected areas. The domains were chosen in a way to avoid any mixing between garnet and phengite generations, and excluding mineral inclusions.

4.2 | Bulk rock chemistry

Bulk rock chemical analyses were conducted at the Institute of Geological Sciences, University of Bern (Switzerland). Rock samples were crushed by using a stainless steel mortar and pestle. They were then dry milled in an agate ring mill for 30 min. Each run of sample was bracketed by sand quartz milling for 15 min in order to clean the receptacle. Nanoparticulate pressed powder pellets (PPPs) were created following the procedure described in Peters and Pettke (2017) by using microcrystalline cellulose as a binder. The PPPs were analysed using a laser ablation inductively coupled plasma mass spectrometer (LA-ICP-MS) consisting of a GeoLas-Pro 193 nm ArF Excimer laser system (Lambda Physik) in combination with an ELAN DRC-e quadrupole mass spectrometer (Perkin Elmer). Calibration was based on GSD-1G (Jochum et al., 2011) reference material. BRP-1 (Cotta et al., 2007) was used as a secondary standard (Peters & Pettke, 2017) and accuracy was always better than 10%. PPPs were measured in single spot mode with an energy density of 6 J/cm² at a repetition rate of 10 Hz and a beam size of 90 µm. Six analyses were measured for each PPP. The surface area was cleaned by pre-ablation using a larger spot size and for each analysis the signal was integrated over a 50–60 s. Data reduction was performed offline by using the software SILLS (Guillong et al., 2008) and limits of detection were calculated after Pettke et al. (2012). Internal standardization was done by assuming (a) a fixed total of 100 wt% for major and trace elements, minus volatiles determined by loss on ignition; (b) all Fe is present as Fe₂O₃.

4.3 | Trace element analysis

LA-ICP-MS mineral analyses were conducted using two different instruments at the Institute of Geological Sciences,

University of Bern (Switzerland). Part of the punctual analyses was achieved with a GeoLas-Pro 193 nm ArF Excimer laser system (Lambda Physik) in combination with an ELAN DRC-e quadrupole mass spectrometer (Perkin Elmer). Mapping and the second part of mineral punctual analyses were conducted using a Resonetics RESolutionSE 193nm excimer laser system equipped with a S-155 large volume constant geometry chamber (Laurin Technic). The laser system was coupled to an Agilent 7900 quadrupole ICP-MS instrument. Calibration was performed using reference material GSD-1G (Jochum et al., 2005), whereas NIST 612 (Jochum et al., 2011) was used as secondary standard with three ablation spots each. Data were reduced offline employing SILLS (Guillong et al., 2008) and Iolite v7.08 (Paton et al., 2011). Drift was linearly corrected by bracketing the unknowns with the calibration material and LODs were calculated according to Pettke et al. (2012). Beam diameter was adapted according to grain size.

On the Resonetics RESolutionSE 193nm excimer laser system garnet was ablated at 5 Hz and a fluence of 4 J/cm², during post-acquisition quantification processing ²⁷Al was used as internal standard (11 wt%). On the GeoLas-Pro 193 nm ArF Excimer laser system garnet was ablated at 10 Hz and 9 J/cm², during post-acquisition quantification processing internal standardization was done by summing the measured mass fraction of element oxides to 100 wt% assuming garnet as anhydrous phase. Accuracy on secondary standard was better than 10% for all elements.

Rutile was ablated on the GeoLas-Pro 193nm ArF Excimer laser system at 8 Hz and a fluence of 6 J/cm². During post-acquisition quantification processing ⁴⁹Ti was used as internal standard (59.94 wt%). Accuracy on secondary standard NIST 612 was within 10% for all elements and 20% for Zr.

Garnet trace element maps were acquired on standard 30 µm thick polished thin sections using a comparable protocol to that described in Raimondo et al. (2017) and George et al. (2018). The laser was operated at a repetition rate of 10 Hz, a fluence of 7 J/cm² and a continuous scan speed of 10 µm/s, using a beam diameter of 12 or 16 µm. Line spacing was adapted accordingly to the beam diameter. A total of

31 elements were measured for a total sweep time of 0.46 s, where dwell times were 0.01 s for most masses with the exception of light REE (LREE; from La to Gd = 0.02 s), and Hf (0.03 s). During post-acquisition quantification processing ^{27}Al was used as internal standard (11 wt%). Image processing was performed using XMapTools 3.2.1.

4.4 | Lu–Hf geochronology

For Lu–Hf garnet geochronology, bulk rock powders and mineral separates were prepared at the Institute of Geological Sciences, University of Bern (Switzerland). High quality mineral separates were obtained by applying high voltage discharge on rock samples using SelFrag Lab system. Mesh size allowing specific mineral fraction separation was adapted for each sample. The fragmented material was then dried in an oven at 60°C. Multi-grain garnet separates were carefully prepared under a binocular microscope. Whole-rock powders were made by crushing rock samples in a stainless steel mortar and pestle. Samples were then dry milled in an agate ring mill for 30 min. Each sample run was bracketed by sandquartz milling for 15 min in order to clean the receptacle. Chemical digestion and isotopic analyses were conducted at the Pacific Centre for Isotope and Geochemical Research, University of British Columbia. Garnet separates and whole-rock powders were weighed in screw-top PFA beakers following admixing of a ^{176}Lu – ^{180}Hf isotope tracer. Garnet dissolution was done by cycling HF – HNO_3 – HClO_4 and HCl with intermittent solution dry-down. This method dissolves garnet without co-dissolving zircon (Lagos et al., 2007; Scherer et al., 2000). Whole-rock powders were dissolved in HF – HNO_3 kept at 180°C for 5 days in Parr® high- P digestion vessels. Solutions containing dissolved mineral and whole-rock chemical matrices were loaded onto pre-conditioned 12 ml Teflon® columns containing Eichrom® Ln-spec resin and Lu and Hf were isolated and purified using the methods of Münker et al. (2001). The Lu and Hf isotope measurements were carried out using a Nu Instruments *Plasma I* multi-collector ICPMS. The $^{176}\text{Lu}/^{175}\text{Lu}$ of unknowns was determined using ^{176}Yb interference corrections on the basis of a linear correlation between $\ln(^{176}\text{Yb}/^{171}\text{Yb})$ – $\ln(^{174}\text{Yb}/^{171}\text{Yb})$, which was determined by replicate Yb standard measurements (Blichert-Toft et al., 2002). Hafnium mass bias was corrected assuming the exponential law and applying $^{179}\text{Hf}/^{177}\text{Hf} = 0.7325$. All Hf isotope ratios are reported relative to JMC-475 ($^{176}\text{Hf}/^{177}\text{Hf} = 0.282163$; Blichert-Toft et al., 1997), for which long-term external reproducibility is 0.3 eHf at the concentration typical of garnet samples. The external reproducibility (2 SD) of the $^{176}\text{Hf}/^{177}\text{Hf}$ of unknowns was estimated on the basis of the external reproducibility for ATI-475—and in-house Hf reference solution that is isotopically identical to, and prepared from the same metal ingots as, JMC-475

($^{176}\text{Hf}/^{177}\text{Hf} = 0.282160$; Blichert-Toft et al., 1997), which was measured at concentrations that bracketed those of the unknowns (Bizzarro et al., 2003). A realistic blanket uncertainty of 0.25% on $^{176}\text{Lu}/^{177}\text{Hf}$ was assumed. Total procedural Hf blanks during the course of the analytical session were 12 pg or lower. Isochron regressions and age calculations were done using Isoplot version 3.27 (Ludwig, 2003), applying $1.867 \times 10^{-11} \text{ year}^{-1}$ for $\lambda^{176}\text{Lu}$ (Scherer et al., 2001; Söderlund et al., 2004). All uncertainties are reported at the 2 SD level.

4.5 | Phase equilibrium modelling

The P – T conditions for garnet formation were determined using the program GrtMod 1.6.2 (Lanari et al., 2017) based on an iterative simulation of garnet growth and resorption. Starting with the bulk rock composition, each stage of garnet growth is modelled for determining P – T and the amount of fractionated garnet in order to match the observed garnet composition. The modelled garnet composition is obtained by minimizing the Gibbs energy. The input parameters of GrtMod are (a) the bulk rock composition, (b) the amount of water available in the system, (c) the oxidation state of the system, and (d) the number of growth stages each of them having a specific garnet composition. The optimization routine produces for each growth stage ‘optimal’ P – T conditions and an uncertainty envelope by minimizing the differences between the measured and the modelled garnet compositions while optimizing the reactive bulk composition, which depends on how much garnet is fractionated (i.e. subtracted from the bulk rock composition) or resorbed (i.e. re-injected in the reactive bulk composition). A software option can be used to prevent resorption in case of isolated growth zones, or clear the absence of resorption textures in the sample. Gibbs energy minimizations were performed using Theriak-Domino (De Capitani & Brown, 1987; De Capitani & Petrakakis, 2010) and the internally consistent thermodynamic data set of Holland and Powell (1998) and subsequent updates gathered in tc55 (see below). The chemical system was restricted to SiO_2 – TiO_2 – Al_2O_3 – FeO – MnO – MgO – CaO – Na_2O – K_2O – H_2O – O_2 . The following solid solution models were used: feldspar (Holland & Powell, 2003); spinel, biotite (White et al., 2007); epidote, cordierite, talc, staurolite, chlorite, carpholite, and garnet (Holland & Powell, 1998); clinopyroxene (Green et al., 2007); chloritoid (White et al., 2000); white mica (Coggon & Holland, 2002); amphibole (Diener et al., 2007). Water saturation was applied to all models and this assumption is also justified for the retrograde stages as the breakdown of lawsonite before garnet growth liberates a significant amount of water. Excess oxygen of 0.005 mol was used for sample Z16TB32 to reproduce the observed mode of epidote. These values were adjusted based on T – X diagrams.

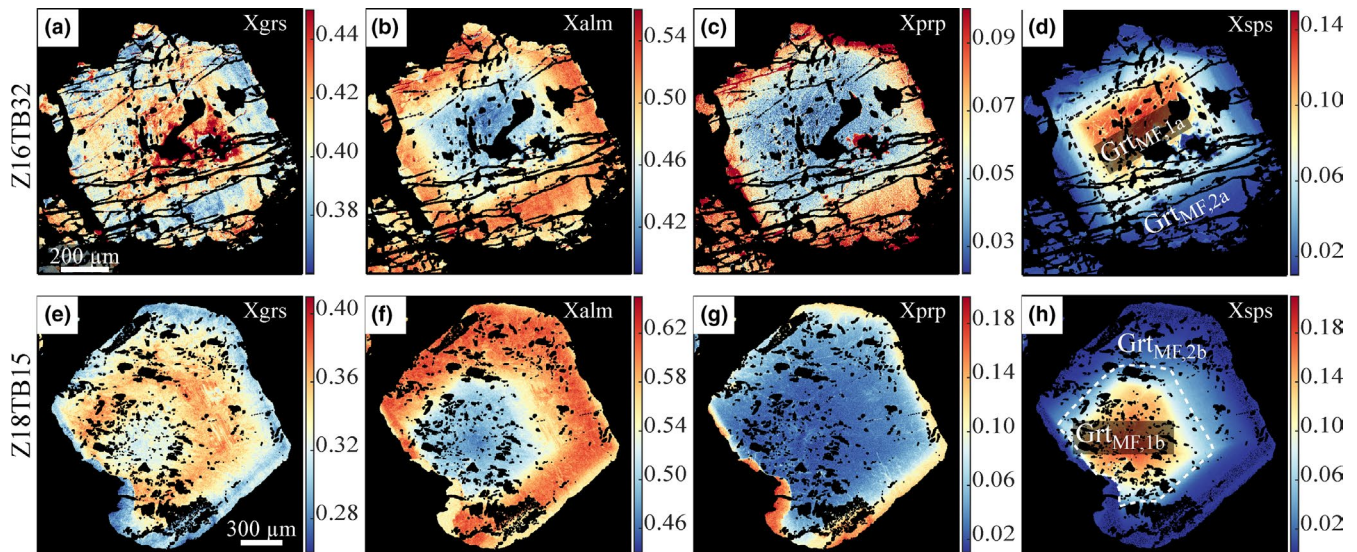


FIGURE 3 End-member compositional maps of garnet generated using XMapTools (Lanari et al., 2014, 2019) from the mafic granofels sample Z16TB32 (a–d) and sample Z18TB15 (e, f)

For each simulated stage, GrtMod produces the following outputs: (a) the reactive bulk composition, (b) the ‘optimal’ P – T conditions, (c) the modelled garnet composition and (d) the volume of garnet fractionated and resorbed for the actual stage and each of the previous garnet generations. The robustness of the results produced by GrtMod was systematically evaluated as follow: (a) The volume of previous garnet fractionated was compared to observations on each sample; (b) the quality of the solution (i.e. the difference between observed and modelled garnet composition) was monitored using the tolerance factor (TF) with a maximal threshold of 0.05 (e.g. Lanari et al., 2017); and (c) the coexisting minerals predicted to be stable for the same reactive bulk composition were simulated using Theriak-Domino at P – T condition of garnet equilibrium. The simulated phase assemblage was compared with mineral inclusions observed in garnet.

Two independent methods were applied for white mica thermobarometry. (a) Isopleth intersection, considering Si in atoms per formula unit (a.p.f.u.) and X_{Mg} ($X_{Mg} = Mg/(Mg+Fe)$), obtained by forward modelling with the software Theriak-Domino. White mica fractionation was not considered as it has no major effect on the modelled compositions (Airaghi et al., 2017). For each compositional group of average X_{Mg} and Si a.p.f.u. values, P – T estimates are presented as an ‘optimal’ solution together with an uncertainty envelope obtained using the method of Lanari et al. (2017) adapted here for K-white mica. (b) Inverse models based on the method and solid-solution model of Dubacq et al. (2010) were calculated and plotted using the program ChlMicaEqui (Lanari, 2012). For a given potassic white mica composition in equilibrium with quartz and water, the equilibrium conditions are represented by a line in P – T space along which the interlayer water content varies (Dubacq et al., 2010).

The P – T conditions of chloritoid were determined using X_{Mg} ($X_{Mg} = Mg/(Mg+Fe)$) isopleths modelled with Theriak-Domino.

4.6 | Zr-in-rutile thermometry

Calculations were performed using the ‘combined model’ of Kohn (2020) assuming growth at the maximum pressure determined for our samples (2.65 GPa; see below), thus providing maximum temperature estimates. The sources of uncertainty to be propagated onto Zr-in-rutile temperature estimates are the effect of analytical uncertainty of Zr measurement and uncertainty inherent in the calibration ($\sim 2\%$ for 500–650°C, Kohn 2020). The Zr analytical error was estimated to be $\pm 20\%$, which corresponds to an uncertainty of ± 15 – 20°C for the 500–650°C temperature range. Combined with the internal error of the calibration of Kohn (2020), a total uncertainty of $\pm 35^\circ\text{C}$ is estimated for the same temperature range. The presence of quartz and zircon in the schists ensures an activity of SiO_2 and ZrO_2 of unity for a robust application of this thermometer.

5 | RESULTS

5.1 | Mineral texture and chemistry

5.1.1 | Garnet chemistry

Major and trace element compositional maps of garnet from the mafic granofels, mafic schist, and Cld-schist are shown in Figures 3–6 and in Figures S1–S4. Pixels were selected

in specified regions-of-interest to reflect the main chemical zones observed in garnet chemical maps (Figures 3d,h, 4c,g, and 5c). The average composition of each growth zone is reported in Tables S1 and S2.

Mafic granofels samples Z16TB32 and Z18TB15

Garnet composition ranges between Grs₃₀ and Grs₅₀ with pyrope content below Prp₂₀. Garnet cores are systematically enriched in Mn up to Sps₂₀ in sample Z18TB15 and Sps₁₅ in Z16TB32. In sample Z16TB32, Ca and Mn decrease from core (Grt_{MF,1a}) to rim (Grt_{MF,2a}) following a bell-shaped profile that is mirrored by Fe and Mg (Figure 3a–d). Irregular Ca enrichment is observed along rims (Figure 3a). In sample Z18TB15, Mn decreases from core (Grt_{MF,1b}) to rim (Grt_{MF,2b}) following a bell-shaped profile associated with Fe increase (Figure 3b,h). The distribution of Ca is more complex, with a Ca-poor core, a Ca-rich mantle, and a Ca-poor rim (Figure 3e). The numerous garnet grains investigated locally show a discontinuous 0.2–0.3 mm rim that is enriched in Mg and depleted in Mn, Fe, and Ca (Figure 3e–h). In sample Z18TB15, garnet shows strong concentric trace element zoning, with maximum heavy REE (HREE) concentrations in cores decreasing towards the rim and local patchy Y+HREE enrichment in mantle (Figure 6a and Figure S1). The discontinuous rim defined by major elements (Figure 3e–h) is

enriched in Co, Zn, Y, Eu, middle REE (MREE), and depleted in Ti, V, Cr, and HREE.

Mafic schist samples Z16TB11 and Z16TB28

The small garnet grains in the matrix exhibit two chemically distinct zones with a sharp and irregular boundary that is less than 1 μm wide (Figures 4a–d and 7a). The cores (Grt_{MS,1}) are enriched in Fe, Mg, and Mn, and depleted in Ca (Alm_{64–68}Prp_{23–27}Grs_{13–15}Sps_{1–3}) compared to the rims. Mn enrichment domains are observed along the core–rim boundary (Figure 4d). The euhedral rims are zoned, with an inner part (Grt_{MS,2}) that is distinctly richer in Ca and poorer in Mg (Alm_{54–59}Prp_{5–8}Grs_{32–36}Sps_{1–2}) than the outer part (Grt_{MS,3}, Alm_{54–59}Prp_{19–22}Grs_{21–25}Sps_{1–2}). Larger garnet porphyroblasts (Figures 4e–l and 7b) are characterized by smooth concentric zoning, with Mg concentrations increasing and Fe and Mn decreasing towards the rim. Ca zoning is locally more patchy and inhomogeneous in sample Z16TB11. The chemical composition of these porphyroblasts (Alm_{54–62}Prp_{8–22}Grs_{24–32}Sps_{1–9}) is similar to that of the rim of matrix garnet, except that Mn core concentrations are slightly higher. In sample Z16TB11, garnet porphyroblasts display only minor trace element zoning (Figure 6b and Figure S2). Euhedral, oscillatory zoning at the 50 μm scale is visible in Cr, Y, and HREE, and sector zoning is

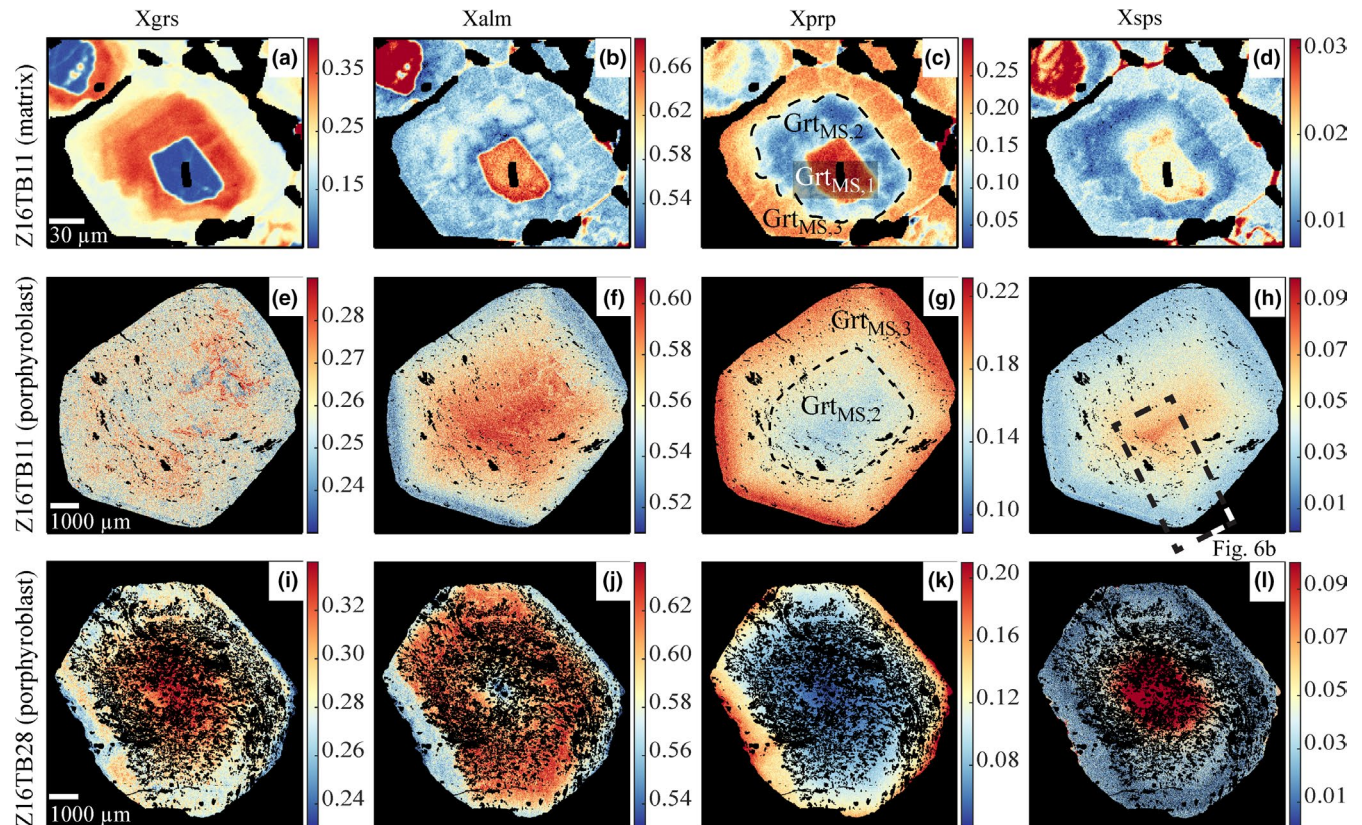


FIGURE 4 End-member compositional maps of small garnet crystals from the mafic schist sample Z16TB11 (a–d), and garnet porphyroblasts from sample Z16TB11 (e–h) and sample Z16TB28 (i–l)

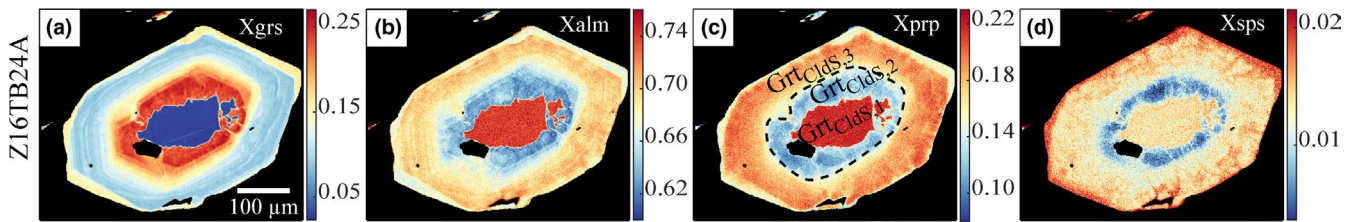


FIGURE 5 End-member compositional maps of garnets from the Cld-schist sample Z16TB24A

observed for Y and REE within garnet mantle and rim. REE patterns (Figure 8a) show an increase from LREE to MREE, and a flat HREE pattern. The REE pattern of garnet porphyroblasts from sample Z16TB28 (Figure 8c) shows an increase from MREE to HREE and a marked decrease in Gd_N/Yb_N (N = chondrite normalized) from core to rim. For the small garnet grains that display sharp chemical zoning in major elements the spatial resolution of the ablation spot is insufficient to resolve all the domains. REE normalized patterns (Figure 8b) show comparable HREE composition between core and rim, with a Gd_N/Yb_N relatively close to unity. The cores are generally richer in MREE and have a marked negative Eu anomaly ($Eu/Eu^* = Eu_N/(Sm_N * Gd_N)^{0.5}$) of ~ 0.1 , which is not observed for the rims.

Chloritoid schist sample Z16TB24A

Garnet has two distinct chemical zones (Figures 5 and 7a) separated by a sharp 1 μm contact. The xenomorphic cores ($Grt_{CldS,1}$) are depleted in Ca and enriched in Fe and Mg ($Alm_{72-76}Prp_{20-24}Grs_{3-6}Sps_1$) compared to the rims. The euhedral rims are zoned with decreasing Ca and increasing Mg, Fe, and Mn from the inner ($Grt_{CldS,2}$) to the outer part ($Grt_{CldS,3}$, $Alm_{54-58}Prp_{15-17}Grs_{26-30}Sps_{0.1-0.2}$ to $Alm_{60-68}Prp_{8-12}Grs_{23-27}Sps_{<1}$ respectively). A Ca- and Mn-rich discontinuous domain of up to 15 μm width is observed along the external rim. Trace element mapping (Figure 6c and Figure S3) highlights the strong chemical difference between core and rim, as described for major elements. The core is enriched in V, Cr, Li, Co, Y, and HREE compared to the inner rim. The external rim is depleted in V and Ti, and enriched in Cr, Co, and MREE, compared to the inner rim. On a logarithmic scale, the REE patterns of garnet cores and rims largely overlap (Figure 8d) but the core is characterized by lower Gd_N and a marked Eu negative anomaly of ~ 0.2 , which is not present in the rim.

5.1.2 | K-white mica and chloritoid major element chemistry

The compositional variability of potassic white mica from mafic schists (Z16TB11 and Z16TB28) and the chloritoid schist (Z16TB24A) was investigated. In every sample,

mica flakes are preferentially oriented along the main foliation and microstructural criteria are not sufficient to discriminate different populations. However, the compositional maps reveal variations in the Si a.p.f.u. and X_{Mg} (Figure 9). Pixels selected in specified regions-of-interest on the Si a.p.f.u. and X_{Mg} binary plots were adjusted to reflect the main textural domains and allow for a discrimination of several populations of K-white mica in each sample (Figure 9). Representative average K-white mica compositions were extracted from the elemental maps and are reported in Table S3.

Mafic schist samples Z16TB11 and Z16TB28

Three K-white mica populations were identified in sample Z16TB11 (Figure 9a,c) and five in Z16TB28 (Figure 9d,f). The three mica populations of Z16TB11 ($Ph_{MS,1a}$, $Ph_{MS,2a}$, and $Ph_{MS,3a}$) are phengite and have comparable microtextural location and spatial distribution than the three-first population of Z16TB28 ($Ph_{MS,1b}$, $Ph_{MS,2b}$, and $Ph_{MS,3b}$). The first mica population ($Ph_{MS,1a}$ and $Ph_{MS,1b}$) is located in the central zone of elongated mica flakes stretched along the main foliation. It has on average 3.33 ± 0.02 Si a.p.f.u. and 0.76 ± 0.02 X_{Mg} in sample Z16TB11 and 3.40 ± 0.02 Si a.p.f.u. and 0.79 ± 0.02 X_{Mg} in sample Z16TB28. The second population of mica ($Ph_{MS,2a}$ and $Ph_{MS,2b}$) occurs either on the edge of the first population clusters or as single grains in the foliation. It has on average Si a.p.f.u. of 3.29 ± 0.02 and X_{Mg} of 0.71 ± 0.02 in sample Z16TB11 and Si a.p.f.u. of 3.34 ± 0.02 and X_{Mg} of 0.72 ± 0.02 in sample Z16TB28. The third mica group ($Ph_{MS,3a}$ and $Ph_{MS,3b}$) is found on the edge of the second population clusters or as isolated grains. The Si a.p.f.u. content is 3.22 ± 0.02 and X_{Mg} is 0.66 ± 0.02 in sample Z16TB11 and Si a.p.f.u. of 3.24 ± 0.02 and X_{Mg} of 0.64 ± 0.02 in sample Z16TB28. The last two populations of K-white mica ($Kwm_{MS,1}$ and $Kwm_{MS,2}$) are only observed in Z16TB28 and are characterized by a lower Si a.p.f.u. than the three previous generations (Figure 9d). They are texturally associated with zoisite and paragonite, and form patchy aggregates, which sometimes rim the previously described mica population (Figure 9f). On average, $Kwm_{MS,1}$ Si a.p.f.u. content is 3.11 ± 0.02 and X_{Mg} is 0.78 ± 0.02 , whereas $Kwm_{MS,2}$ Si a.p.f.u. content is 3.12 ± 0.02 and X_{Mg} is 0.67 ± 0.02 .

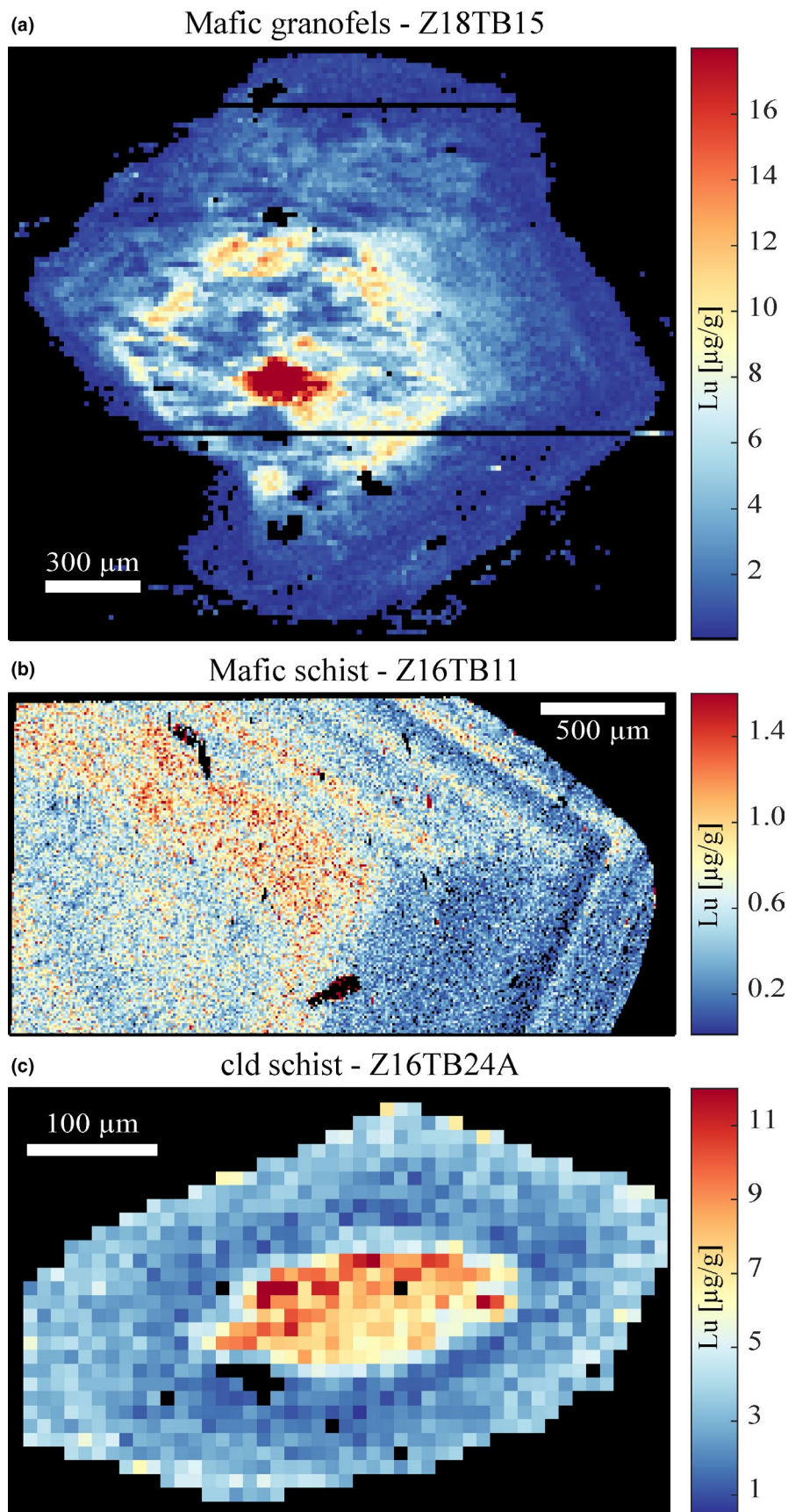


FIGURE 6 Trace element LA-ICP-MS compositional maps of Lu in garnet for the three rock types dated by Lu–Hf. For the mafic schist, Lu distribution in Z16TB11 is interpreted as representative for the dated sample Z16TB28 (see text)

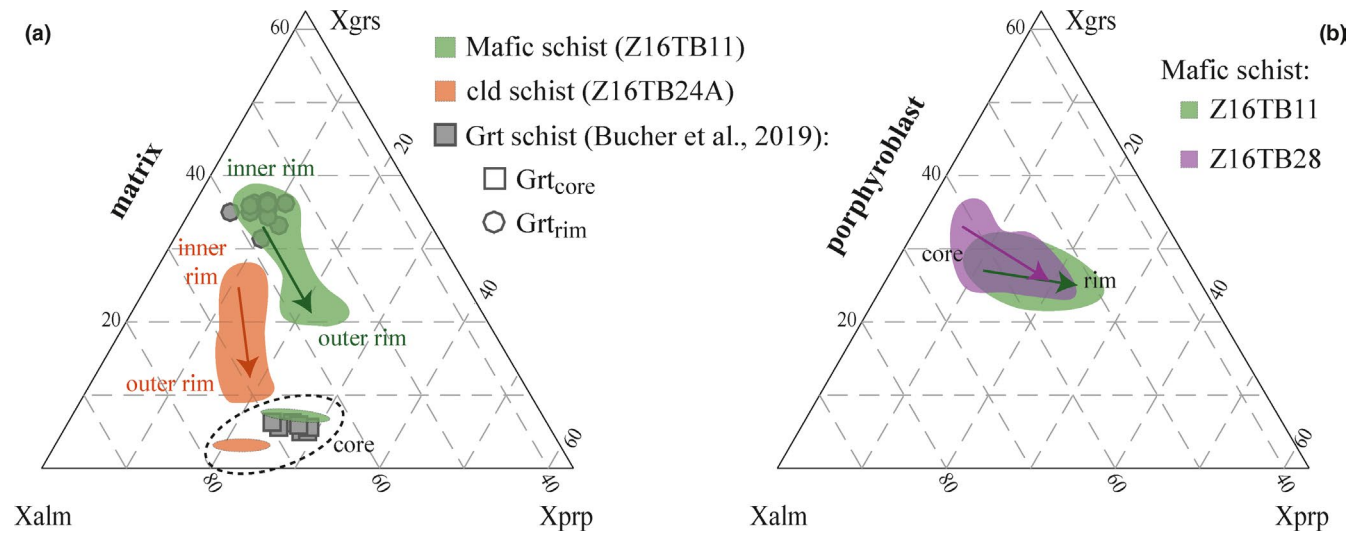


FIGURE 7 Ternary chemical diagram showing the range of garnet composition in $X_{\text{grs}}-X_{\text{alm}}-X_{\text{prp}}$. Range of compositional data was exported from compositional maps (Figure S6). (a) Small garnet grains from the mafic schist (Z16TB11, area drawn on the basis of 30137 pixels) and Cld-schist (Z16TB24A, area drawn on the basis of 111072 pixels), compared with garnet compositions from Bucher et al. (2019). (b) Garnet porphyroblasts from the mafic schist samples Z16TB11 (area drawn on the basis of 516869 pixels) and Z16TB28 (area drawn on the basis of 209512 pixels)

Chloritoid schist sample Z16TB24A

Two populations of K-white mica were identified (Figure 9g). The first population of phengite ($\text{Ph}_{\text{CldS},1}$) is located in the central zone of elongated mica aggregates stretched along the main foliation (Figure 9i). It has on average 3.38 ± 0.02 Si a.p.f.u. and 0.72 ± 0.02 X_{Mg} . The second population of phengite ($\text{Ph}_{\text{CldS},2}$) occurs either on the edge of the first population clusters or as single grains in the foliation. It has on average Si a.p.f.u. of 3.28 ± 0.02 and X_{Mg} of 0.65 ± 0.02 . Chloritoid from the Cld-schist (Table S4) varies in X_{Mg} from 0.25 to 0.38, where higher values are representative of the core and decrease towards the external part of the rim (Figure S5).

5.1.3 | Rutile

Among the studied samples, rutile was found in the mafic and Cld-schist with a grain size varying from a few to 100 μm . All analysed rutile grains were located in the matrix except two grains found as inclusions in garnet (mafic schist Z16TB11 and Z16TB28). Grains for analyses were carefully selected to avoid possible contamination by inclusions and fractures. Analyses (Table S5) were screened for contamination from inclusions based on high Si, Na, Al, Mn, Fe, Y, Sn, Hf, and Zr contents. The average concentration of Zr in rutile (Table 2) within the mafic schist is 36 ± 3 $\mu\text{g/g}$ in sample Z16TB11 and 40 ± 4 $\mu\text{g/g}$ in sample Z16TB28. Rutile inclusions in garnet contain 36 $\mu\text{g/g}$ (sample Z16TB11) and 27 $\mu\text{g/g}$ (sample Z16TB28) Zr. The average Zr concentration in the rutile from Cld-schist (sample Z16TB24A) is 40 ± 5 $\mu\text{g/g}$. The Zr

concentration in rutile is similar within uncertainty among the different lithologies.

5.2 | Thermobarometry

5.2.1 | Thermodynamic modelling

The P - T stability conditions of garnet, phengite, and chloritoid were determined for the investigated samples using the strategy outlined in Section 4.5; results are shown in Figures 9–11 and Table 3. The reactive bulk compositions used are reported in Table S6. After careful identification of representative chemical garnet zones in each sample, average garnet chemical compositions were extracted from the elemental maps (Table S1) from the domains shown in Figures 3d,h, 4c,g, and 5c. Minor chemical variations within the domains for sample Z18TB15, Z16TB24A, and Z16TB11 were not considered.

Mafic granofels sample (Z16TB32)

The garnet core $\text{Grt}_{\text{MF},1a}$ (Figure 3d) is predicted to be stable at $490 \pm 10^\circ\text{C}$ and 1.74 ± 0.05 GPa (TF: 0.011) with 5 vol.% of garnet (Figure 10a). The modelled phase assemblage is $\text{Grt} + \text{Omp} + \text{Chl} + \text{Bt} + \text{Tr} + \text{Lws} + \text{Ttn} + \text{Qz}$. The rim stage $\text{Grt}_{\text{MF},2a}$ (Figure 3d) is modelled at $570 \pm 10^\circ\text{C}$ and 2.52 ± 0.10 GPa (TF: 0.023; 8 vol.% garnet) while preventing resorption of $\text{Grt}_{\text{MF},1a}$ (Figure 10a). The modelled phase assemblage at these conditions is $\text{Grt} + \text{Omp} + \text{Ph} + \text{Lws} + \text{Qz} + \text{Rt}$.

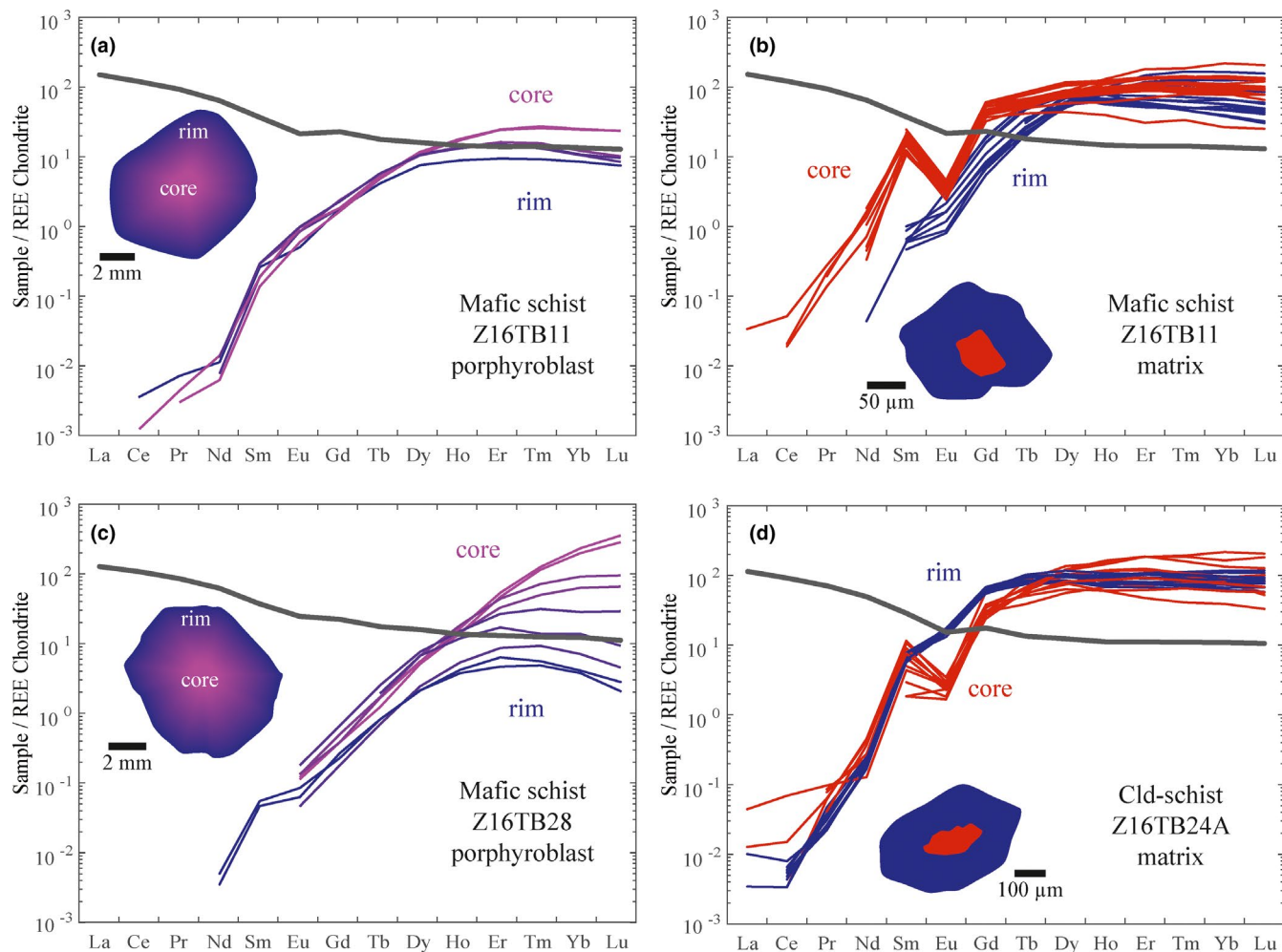


FIGURE 8 REE chondrite-normalized concentrations of bulk rock (black) and garnet (red, violet, and blue) from different samples. Normalization data: Taylor and McLennan (1985)

Mafic granulites sample (Z18TB15)

The garnet core $\text{Grt}_{\text{MF},1\text{b}}$ (Figure 3h) is predicted to be stable at $486 \pm 15^\circ\text{C}$ and 1.81 ± 0.05 GPa (TF: 0.046) with 2 vol.% of garnet (Figure 10b). The modelled phase assemblage is $\text{Grt} + \text{Omp} + \text{Chl} + \text{Bt} + \text{Tr} + \text{Lws} + \text{Qz} + \text{Rt}$. The rim stage $\text{Grt}_{\text{MF},2\text{b}}$ (Figure 3h) is modelled at $590 \pm 15^\circ\text{C}$ and 1.65 ± 0.15 GPa (TF: 0.003) with the formation of 21 vol.% of garnet without requiring resorption of $\text{Grt}_{\text{MF},1\text{b}}$ (Figure 10b). The modelled phase assemblage at these conditions is $\text{Grt} + \text{Omp} + \text{Ms} + \text{Tr} + \text{Zo} + \text{Qz} + \text{Rt}$.

Mafic schists (sample Z16TB11)

The garnet core composition $\text{Grt}_{\text{MS},1}$ (Figure 5c) is modelled at a temperature of $591 \pm 10^\circ\text{C}$ and pressure ranging between 2.65 and 2.85 GPa with optimal solution at 2.65 GPa (TF: 0.039). The modelled phase assemblage is $\text{Grt} + \text{Omp} + \text{Ph} + \text{Gln} + \text{Ky} + \text{Lws} + \text{Qz} + \text{Rt}$ (Figure S7) and contains 19 vol.% of garnet (Figure 10c). Simulation for the inner rim stage $\text{Grt}_{\text{MS},2}$ (Figure 5c) predicts strong resorption of $\text{Grt}_{\text{MS},1}$ (17 vol.%) and leads to the formation of 6 vol.% of garnet $\text{Grt}_{\text{MS},2}$ (Figure 10c) at $553 \pm 10/-50^\circ\text{C}$ and

$1.47 \pm 0.15/-0.35$ GPa (TF: 0.059). The modelled phase assemblage for $\text{Grt}_{\text{MS},2}$ is $\text{Grt} + \text{Chl} + \text{Pg} + \text{Bt} + \text{Gln} + \text{Zo} + \text{Qz} + \text{Rt}$ (Figure S8). The simulation for the external rim $\text{Grt}_{\text{MS},3}$ (Figure 5c) predicts the observed garnet composition at $595 \pm 25/-10^\circ\text{C}$ and $1.80 \pm 0.05/-0.30$ GPa (TF: 0.034). Previously fractionated garnet stages are only slightly affected with 1 vol.% resorption of $\text{Grt}_{\text{MS},1}$, with up to 14 vol.% garnet fractionation of $\text{Grt}_{\text{MS},3}$ (Figure 10c). The modelled phase assemblage is $\text{Grt} + \text{Chl} + \text{Ph} + \text{Pg} + \text{Gln} + \text{Zo} + \text{Qz} + \text{Rt}$ (Figure S9). Intersection of Si a.p.f.u. and X_{Mg} isopleths in phengite (Figure 9b) is observed at $590 \pm 15^\circ\text{C}$ and 2.5 ± 0.1 GPa for $\text{Ph}_{\text{MS},1\text{a}}$. The second population $\text{Ph}_{\text{MS},2\text{a}}$ is modelled at lower pressure and slightly lower temperature of 2.0 ± 0.1 GPa and $550 \pm 10^\circ\text{C}$ respectively. The last group $\text{Ph}_{\text{MS},3\text{a}}$ shows lower pressure conditions (1.6 ± 0.1 GPa) and similar temperature ($560 \pm 10^\circ\text{C}$) compared to $\text{Ph}_{\text{MS},2\text{a}}$. Assuming similar temperature for each population phengite–quartz–water method provides pressure estimates of 2.2 ± 0.1 GPa ($590 \pm 15^\circ\text{C}$, $\text{Ph}_{\text{MS},1\text{a}}$), 1.8 ± 0.1 GPa ($550 \pm 10^\circ\text{C}$, $\text{Ph}_{\text{MS},2\text{a}}$), and 0.5 ± 0.1 GPa ($560 \pm 10^\circ\text{C}$, $\text{Ph}_{\text{MS},3\text{a}}$).

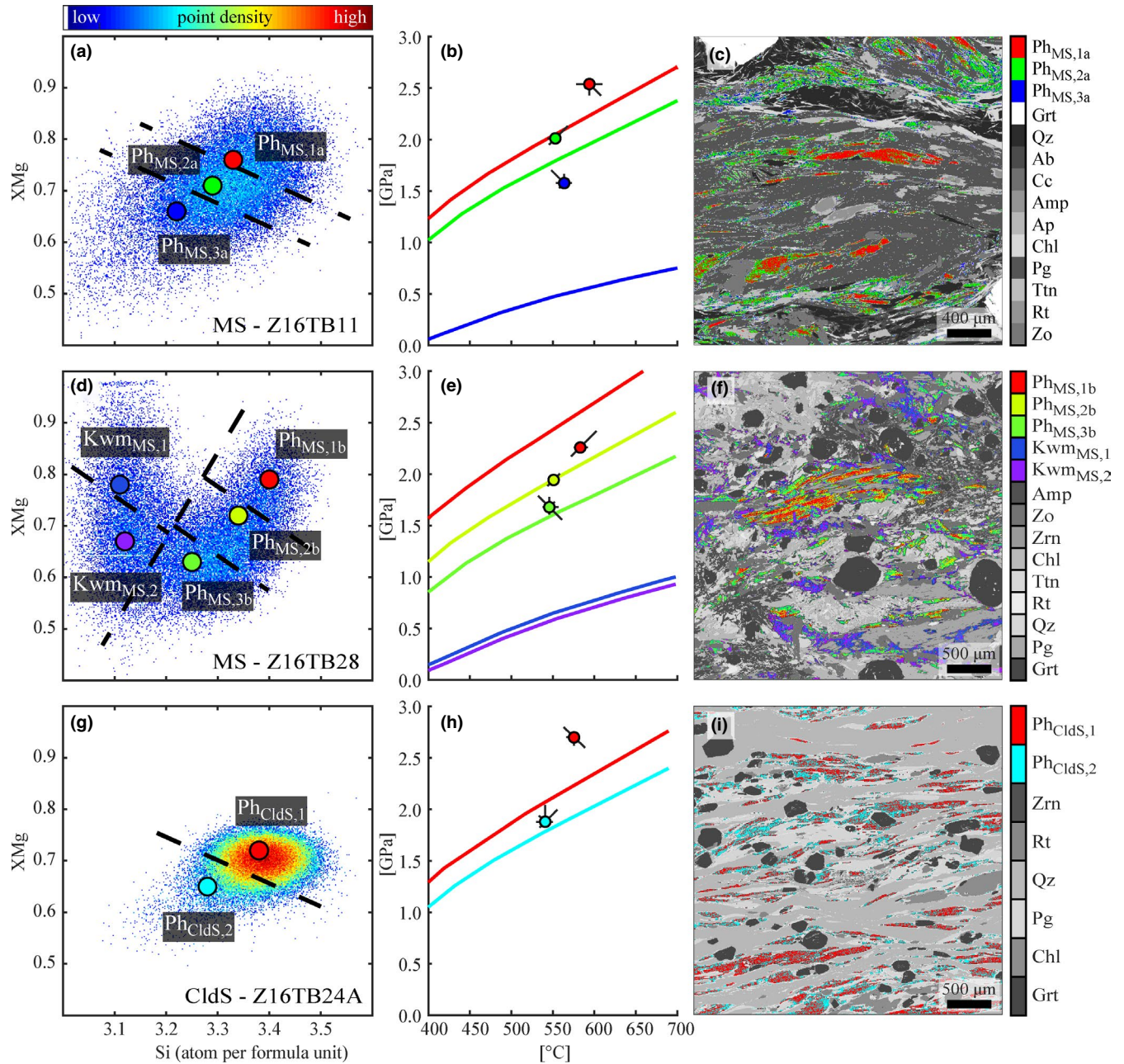


FIGURE 9 K-white mica population discrimination plots. Each row corresponds to a sample: mafic schist Z16TB11 (a–c), mafic schist Z16TB28 (d–f), Cld-schist Z16TB24A (g–i). First column: Si(atom per formula unit)— X_{Mg} binary diagram with colour scale corresponding to point density (log-scale). Spots show the average composition of each K-white mica group used for modelling. Dashed lines represent the boundary between the different compositional groups. Second column: P – T estimates based on mica compositions. The spots represent the intersection between X_{Mg} and Si a.p.f.u. isopleths, whereas lines correspond to phengite–quartz–water equilibria (see text). Third column: spatial relation among K-white mica groups and other phases (in grey)

Mafic schist sample Z16TB28

Si a.p.f.u. and X_{Mg} isopleths of phengite (Figure 9e) intersect at $580 \pm 20/-15^\circ\text{C}$ and $2.3 \pm 0.3/-0.1$ GPa for the first group Ph_{MS,1b}, and $550 \pm 10^\circ\text{C}$ and 2.0 ± 0.1 GPa for the second group Ph_{MS,2b}. The third population Ph_{MS,3b} is modelled at $550 \pm 10^\circ\text{C}$ and 1.7 ± 0.1 GPa, thus at lower pressure conditions compared to Ph_{MS,2b}. Assuming similar temperature for each population, the

phengite–quartz–water method retrieved pressures of 2.6 ± 0.1 GPa ($580 \pm 20^\circ\text{C}$, Ph_{MS,1b}), 2.0 ± 0.1 GPa ($550 \pm 10^\circ\text{C}$, Ph_{MS,2b}), and 1.6 ± 0.1 GPa ($550 \pm 10^\circ\text{C}$, Ph_{MS,3b}). Populations Kwm_{MS,1} and Kwm_{MS,2} can only be modelled using the method of Dubacq et al. (2010)—there is no intersection of the isopleths in the P – T space—and both indicate much lower pressure conditions below 1 GPa (Figure 9e).

TABLE 2 Thermometric estimates determined by Zr-in-rutile thermometry (Kohn, 2020)

Lithology	Sample	N	Zr-rutile (µg/g)				Pressure (GPa)	Zr-in-Rutile T (°C)			
			Min	Max	Average	SD		Min	Max	Average	Error
Matrix											
Mafic schist	Z16TB11	16	31	40	36	3	2.65	524	540	534	35
Mafic schist	Z16TB28	25	33	49	40	4	2.65	528	555	541	35
Cld-schist	Z16TB24A	24	31	49	40	5	2.65	524	555	541	35
Inclusion in garnet											
Mafic schist	Z16TB11	1	-	-	36	-	2.65	-	-	534	35
Mafic schist	Z16TB28	1	-	-	27	-	2.65	-	-	515	35

Number of grain analysed (N), minimum (Min), maximum (Max), standard deviation (SD).

Chloritoid schist sample Z16TB24A

The garnet core composition $\text{Grt}_{\text{CldS},1}$ (Figure 4c) is predicted to be stable at $576 \pm 10^\circ\text{C}$ and $2.70 + 0.30/-0.10$ GPa (TF: 0.046). The modelled phase assemblage is $\text{Grt}+\text{Omp}+\text{Car}+\text{Ph}+\text{Ky}+\text{Lws}+\text{Qz}+\text{Rt}$, and contains 11 vol.% of garnet (Figure 10d). Simulation of the inner rim stage $\text{Grt}_{\text{CldS},2}$ (Figure 4c) indicates resorption of previous stage $\text{Grt}_{\text{CldS},1}$ (7 vol.%) and the formation of 6 vol.% of garnet $\text{Grt}_{\text{CldS},2}$ (Figure 10d) at $573^\circ\text{C} + 10/-30^\circ\text{C}$ and 1.49 ± 0.10 GPa (TF: 0.048). The modelled phase assemblage is $\text{Grt}+\text{Chl}+\text{Ph}+\text{Pg}+\text{Bt}+\text{Qz}+\text{Rt}$. The simulation for the external rim $\text{Grt}_{\text{CldS},3}$ (Figure 4c) is modelled at $585 \pm 10^\circ\text{C}$ and $2.05 + 0.05/-0.40$ GPa (TF: 0.049). Previously formed garnet growth zones are slightly affected with 2 vol% resorption of $\text{Grt}_{\text{CldS},1}$ and 3 vol.% of $\text{Grt}_{\text{CldS},2}$ respectively (Figure 10d). The associated paragenesis is $\text{Grt}+\text{Ctd}+\text{Ph}+\text{Pg}+\text{Gln}+\text{Qz}+\text{Rt}$. The intersection of Si a.p.f.u. and X_{Mg} phengite isopleths (Figure 9h) provides conditions of $570 \pm 10^\circ\text{C}$ and 2.7 ± 0.1 GPa for $\text{Ph}_{\text{CldS},1}$ and $540 \pm 10^\circ\text{C}$ and 1.9 ± 0.1 GPa for $\text{Ph}_{\text{CldS},2}$. Assuming a similar temperature for each population the phengite-quartz-water method indicates pressure of 2.3 ± 0.1 GPa ($570 \pm 10^\circ\text{C}$, $\text{Ph}_{\text{CldS},1}$) and 1.9 ± 0.1 GPa ($540 \pm 10^\circ\text{C}$, $\text{Ph}_{\text{CldS},2}$). The X_{Mg} isopleth of chloritoid (Figure 11a) is compatible with chloritoid core formation at $580 \pm 10^\circ\text{C}$ between 1.8 and 2.5 ± 0.1 GPa. The external rim is modelled between 1.5 and 2.5 ± 0.1 GPa for a temperature of $550 \pm 10^\circ\text{C}$.

5.3 | Zr-in-rutile thermometry

Zr-in-rutile temperatures were calculated assuming a pressure of 2.65 GPa to provide an independent constraint on the temperature (Table 2). Averaged temperature estimates for rutile found in the mafic schist are $534 \pm 35^\circ\text{C}$ for sample

Z16TB11 and $541 \pm 35^\circ\text{C}$ for sample Z16TB28. The two rutile grains found as inclusions in garnet gave $534 \pm 35^\circ\text{C}$ (sample Z16TB11) and $515 \pm 35^\circ\text{C}$ (sample Z16TB28) respectively. In the Cld-schist (sample Z16TB24A) the average estimated temperature is $541 \pm 35^\circ\text{C}$. These results witness homogeneous temperature of crystallization among the lithologies of TGU. The dependence of the Zr-in-rutile thermometer on pressure implies that these are maximum temperature for rutile formation.

5.4 | Garnet Lu-Hf chronology

The garnet crystals observed in the different lithologies of the TGU present high variability in size, composition, and texture. Key samples were selected (Figure S10) to correlate the garnet size fraction with garnet compositional zoning. In the mafic granofels sample Z18TB15, the compositional zoning in garnet is comparable among all the grains at a sample scale and two garnet fractions were selected, the first one having a diameter between 0.25 and 0.50 mm and the second comprising larger grains (>0.50 mm). Four garnet fractions, comprising two from either size fraction, yielded a Lu-Hf age of 50.3 ± 0.3 Ma (mean square weighted deviation [MSWD] = 1.5; Figure 12a).

Two garnet types were targeted for the schists (Z16TB28 and Z16TB24A). In the mafic schist Z16TB28, garnet porphyroblasts do not show complex textures and have a restricted compositional range. In sample Z16TB28, garnet grains with diameter over 2 mm were picked. The material yielded a Lu-Hf age of 49.6 ± 0.8 Ma (MSWD = 0.34; $n = 3$; Figure 12b). Small garnets with complex and sharp chemical zoning are present in the Cld-schist Z16TB24A and in mafic schist. This garnet type was picked in the Cld-schist because this is the only garnet type observed

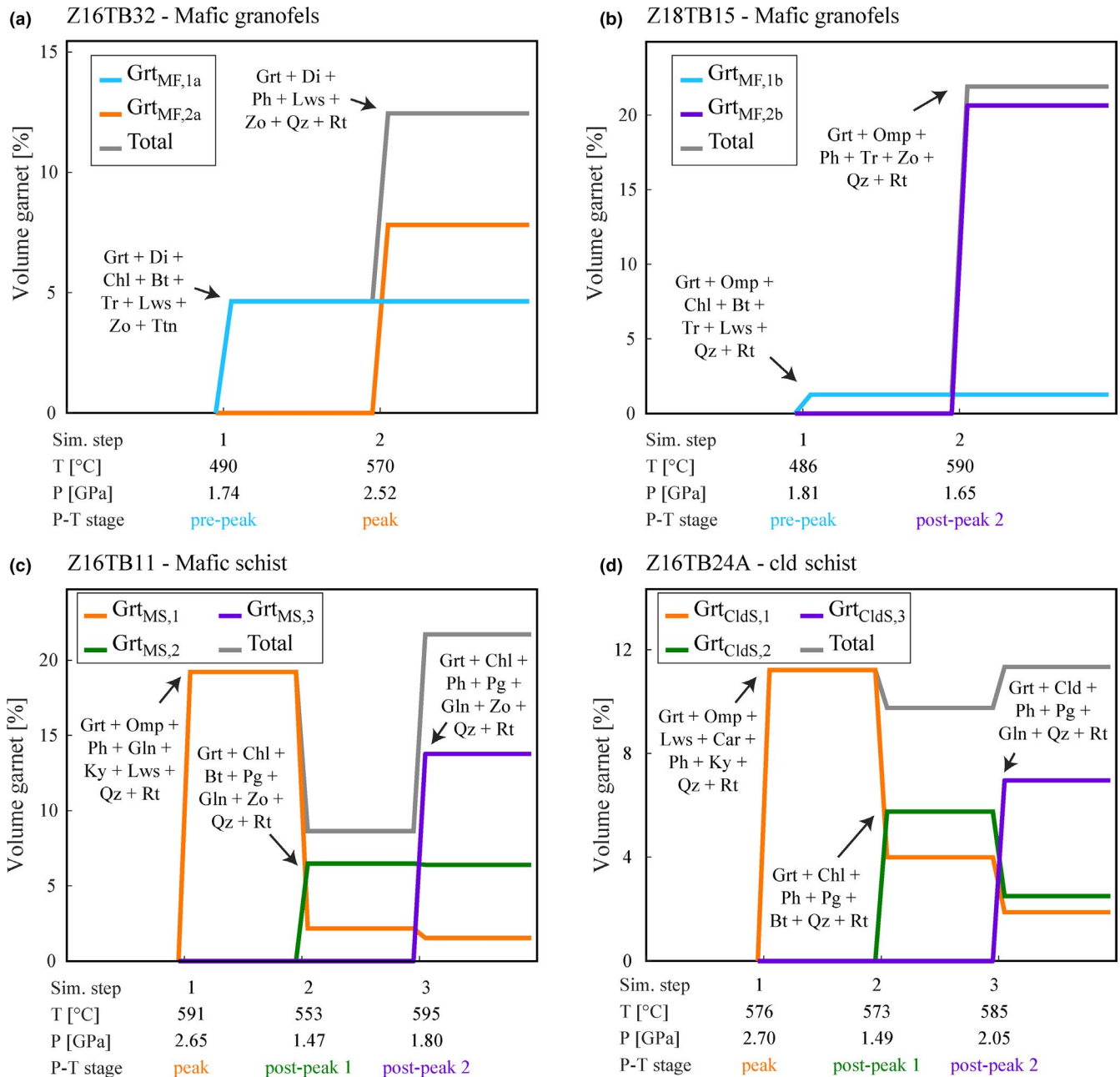


FIGURE 10 Garnet growth/resorption history for different samples obtained using GrtMod. The evolution of volume fraction (vol.%) of each growth zone are plotted for each step. The curve labelled ‘Total’ shows the evolution of the total amount of garnet in the system. The P – T conditions of each step are the optimal conditions determined by the model. The modelled mineral assemblages are indicated for each garnet growth step

in this lithology. Two garnet fractions were differentiated in sample Z16TB24A: the first fraction contains garnet grains with diameter between 250 and 500 μm and the second garnet grains with diameter $>500 \mu\text{m}$. The sample yielded a Lu–Hf age of $48.8 \pm 0.1 \text{ Ma}$ (MSWD = 1.4; $n = 4$; Figure 12c). All samples yielded Lu–Hf dates with uncertainties of 1%RSD or better (Table 4). The relatively high $^{176}\text{Lu}/^{177}\text{Hf}$ for Z18TB15 (>4) and Z16TB24A (>13) returned extremely good precision of $\pm 0.3\%$ RSD and $\pm 0.2\%$ RSD respectively.

6 | DISCUSSION

6.1 | Mineral sequence and P – T path of the TGU

The multiple, independent P – T constraints obtained from the TGU samples can be divided into peak- P metamorphic conditions and conditions that represent earlier (pre-peak) and later (post-peak) metamorphism (Figures 10 and 11a, Table 3). These distinct metamorphic stages were determined

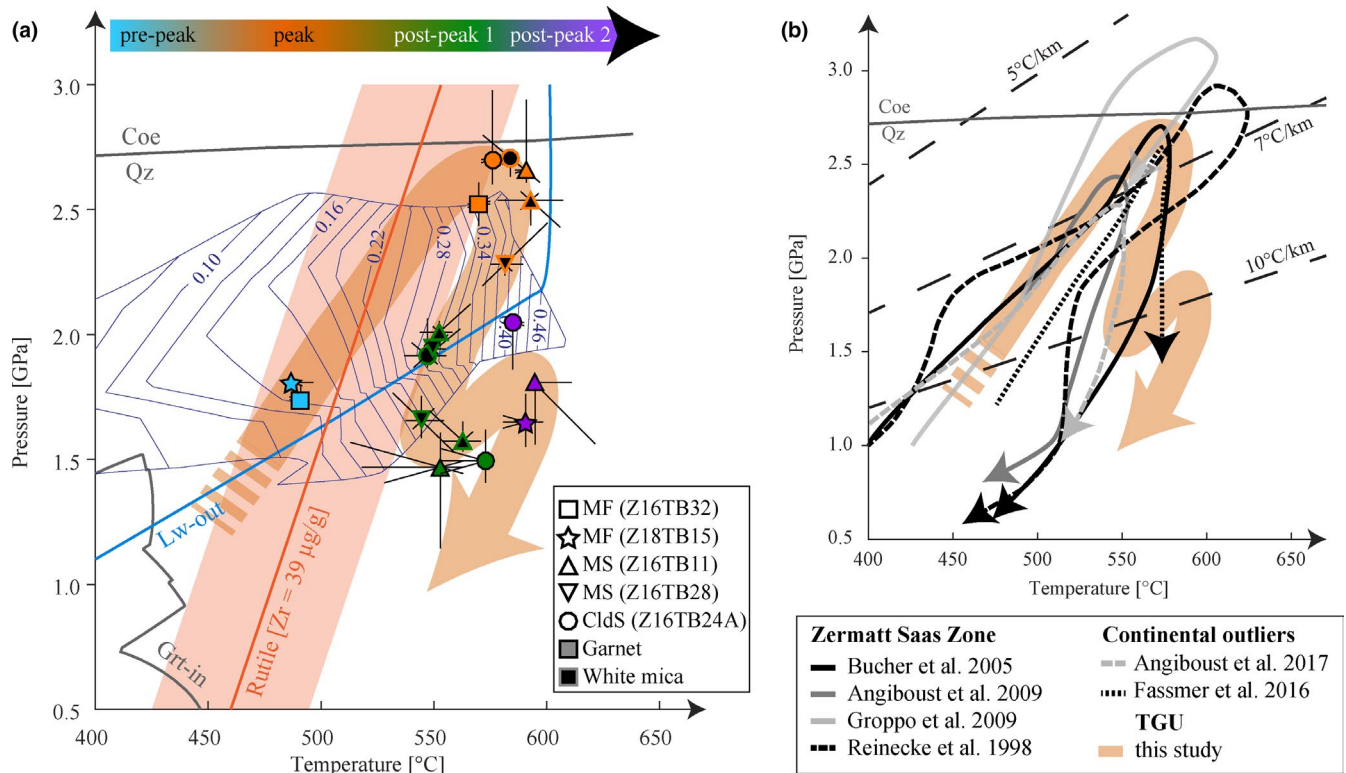


FIGURE 11 Pressure–temperature path of the TGU and the ZSZ tectonic unit. (a) Synthetic P – T diagram compiling the thermobarometric results from the four investigated samples. Abbreviation for the lithologies are: MF—mafic granofels, MS—mafic schist, CldS—Cld-schist. Theriak-Domino results: (i) isopleths shown in blue are X_{Mg} in chloritoid (Cld-schist, Z16TB24A); (ii) The Grt-in curve (mafic granofels, Z18TB15) is shown in grey. (b) P – T diagram showing a compilation of published P – T paths from other localities in the ZSZ tectonic unit and published P – T path from continental outlier in the vicinity of the TGU

according to P – T estimates simulated in the different mineral growth zones described previously.

Pre-peak metamorphic conditions are inferred from the numerical simulation of garnet for the mafic granofels samples Z16TB32 and Z18TB15. Garnet cores Grt_{MF,1a} and Grt_{MF,1b} record the earliest growth of garnet at P – T conditions of $490 \pm 15^\circ\text{C}$ and 1.75 ± 0.05 GPa (Figure 11a). Equilibrium phase diagram calculations using pristine bulk rock compositions show that the garnet-in curve is at a temperature that is $\sim 50^\circ\text{C}$ lower than these values (Figure 11a). This discrepancy could be explained either by overstepping to provide the driving force necessary for porphyroblast nucleation and growth (Spear, 2017; Spear & Pattison, 2017; Spear & Wolfe, 2019; Wolfe & Spear, 2018) or off-centre sectioning of garnet. This would have particular impact on the apparent Mn concentration of the garnet nucleus, which is essential to constrain the initiation of the metamorphic growth based on equilibrium thermodynamic modelling (Lanari et al., 2017).

At the mineral scale, fluid–rock interaction commonly leads to fracturing (Angiboust et al., 2011; Giuntoli, Lanari, Burn, et al., 2018; Giuntoli et al., 2018) and pseudomorphic (and commonly non-isochemical) replacement via interface-coupled dissolution and re-precipitation (Konrad-Schmolke et al., 2018; Putnis, 2002), which may overprint previous

chemical information. At the sample scale such fluid–rock interactions could modify the bulk rock composition. These issues are the prime reason why later metasomatism can compromise the reliability of thermodynamic models based on the bulk rock composition (Evans, 2004). In our samples, garnet textures in mafic schists (Figure 4a–d) and Cld-schist (Figure 5), as well as oxygen isotope variations in garnet (Bovay et al., 2021), suggest strong fluid–rock interaction in the post-peak stage. It is thus possible that, of the two garnet generations observed, only the latest stage (Grt_{MS,2}, Grt_{MS,3}, Grt_{CldS,2}, and Grt_{CldS,3}) represents an equilibrium with the bulk rock as measured. Nevertheless, the P – T conditions estimated for peak- P metamorphism ($580 \pm 15^\circ\text{C}$ and 2.65 ± 0.10 GPa) are in remarkable agreement across lithology. These results are consistent with maximum Zr-in-rutile temperature estimates retrieved from homogeneous rutile populations present in different schist samples (Figure 11a). Moreover, K-white mica found in the mafic schist and in the Cld-schist also record similar peak- P metamorphism conditions (Figure 11a), as determined on the basis of two independent methods: (a) the pressure dependency of Si a.p.f.u. content and the temperature dependency of X_{Mg} in phengite (Massonne & Schreyer, 1987), and (b) the quartz–water–phengite barometer of Dubacq et al. (2010). Potassic white

TABLE 3 Results of the thermodynamic modelling (Garnet and Phengite)

Rock type	Mafic granofels		Mafic schist		Mafic schist		Cld-schist			
Sample	Z16TB32	Z18TB15	Z16TB11		Z16TB28		Z16TB24A			
Mineral	garnet	garnet	garnet	phengite	phengite	phengite	garnet	phengite		
Method	GrtMod	GrtMod	GrtMod	Isopleth	Inverse modelling	Isopleth	Inverse modelling	GrtMod	Isopleth	Inverse modelling
Pre-peak										
Population			Grt _{MF,1a}	Grt _{MF,1b}						
<i>T</i> (°C)			490 ± 10	486 ± 15						
<i>P</i> (GPa)			1.74 ± 0.05	1.81 ± 0.05						
Peak										
Population			Grt _{MF,2a}		Grt _{MS,1}	Ph _{MS,1a}	Ph _{MS,1a}	Ph _{MS,1b}	Ph _{MS,1b}	Ph _{CldS,1}
<i>T</i> (°C)			570 ± 10		591 ± 10	590 ± 15	(590 ± 15)*	580 + 20/−15	(580 ± 20)*	570 ± 10
<i>P</i> (GPa)			2.52 ± 0.10		2.65 + 0.20/−0.05	2.5 ± 0.1	2.2 ± 0.1	2.3 + 0.3/−0.1	2.6 ± 0.1	2.70 + 0.30/−0.10
Post-peak 1										
Population					Grt _{MS,2}	Ph _{MS,2a}	Ph _{MS,2a}	Ph _{MS,2b}	Ph _{MS,2b}	Ph _{CldS,2}
<i>T</i> (°C)					553 + 10/−50	550 ± 10	(550 ± 10)*	550 ± 10	(550 ± 10)*	540 ± 10
<i>P</i> (Gpa)					1.47 + 0.15/−0.35	2.0 ± 0.1	1.8 ± 0.1	2.0 ± 0.1	2.0 ± 0.1	1.49 ± 0.10
Population						Ph _{MS,3a}	Ph _{MS,3a}	Ph _{MS,3b}	Ph _{MS,3b}	
<i>T</i> (°C)						560 ± 10	(560 ± 10)*	550 ± 10	(550 ± 10)*	
<i>P</i> (GPa)						1.6 ± 0.1	0.5 ± 0.1	1.7 ± 0.1	1.6 ± 0.1	
Post-peak 2										
Population				Grt _{MF,2b}	Grt _{MS,3}				Grt _{CldS,3}	
<i>T</i> (°C)				590 ± 15	595 + 25/−10				585 ± 10	
<i>P</i> (GPa)				1.65 ± 0.15	1.8 + 0.05/−0.30				2.05 + 0.05/−0.40	

Note: ()* data used for calculation (see text), the different methods are described in the manuscript.

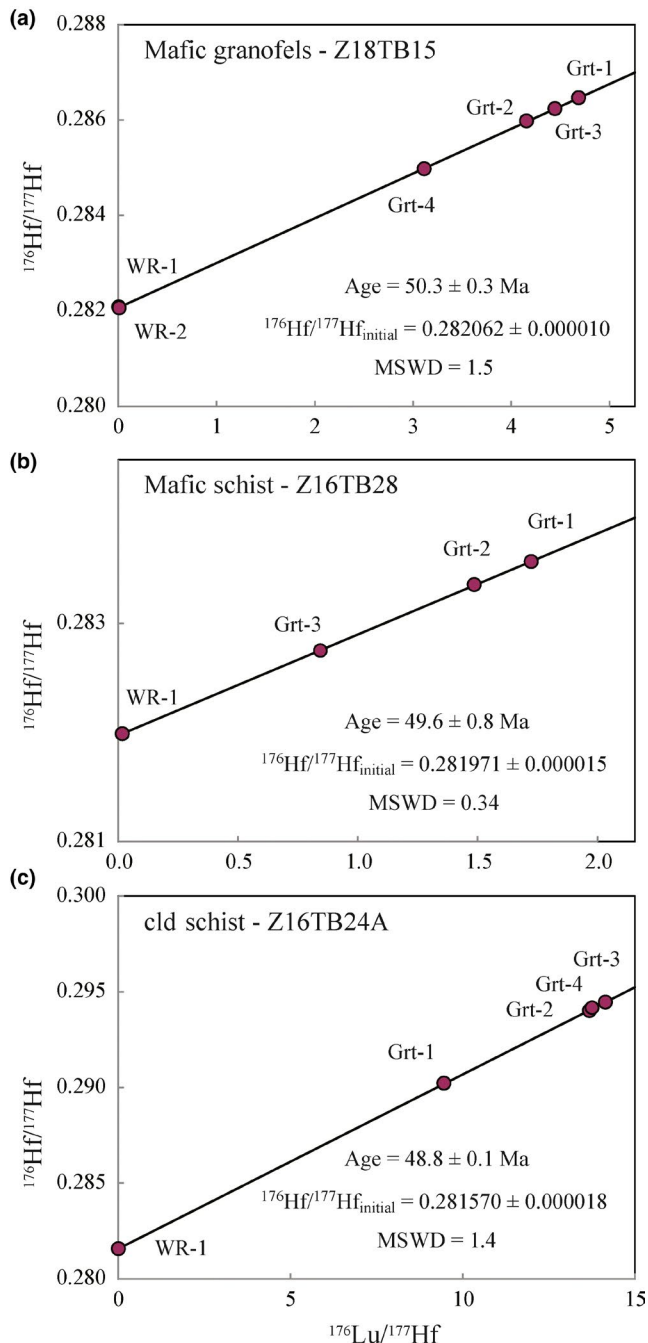


FIGURE 12 Lu–Hf isochrones for the studied samples. (a) Mafic granofels Z18TB15. (b) Mafic schist Z16TB28. (d) Cld-schist Z16TB24A

mica P – T conditions obtained by multi-equilibrium show similar results to the phase diagrams, with variations between samples of less than 0.5 GPa (Figure 9b,e,h) that are within the absolute uncertainty of the method (± 0.25 GPa; Lanari & Duisterhoeft, 2019). The good agreement between P – T estimates from different samples and using different thermobarometric methods—some of them not relying on the reactive bulk composition—indicates minor or insignificant chemical change for major and minor elements by metasomatism and

validates the use of bulk rock chemistry for phase equilibrium modelling in the case of pre-peak and peak- P conditions.

The cores of garnet in Cld-schist Z16TB24A and mafic schist Z16TB11 are attributed to the peak- P stage, and they show similar REE trends (Figure 8b,d) with minor negative Eu anomalies. The absence of plagioclase at these high- P conditions and the lack of evidence for changes in oxidation state exclude these two factors as possible causes for the Eu anomaly (e.g. Ague, 2017). The weak Eu anomaly in peak- P garnet is either inherited from the bulk rock (Gauthiez-Putallaz et al., 2016; Grevel et al., 2010) (Figure 8, Table S7) or due to the presence of lawsonite during prograde metamorphism (Figure 10), as confirmed by modelling and field evidence (Figure 2c). Lawsonite can have a significant positive Eu-anomaly and thus has the capacity to influence Eu budget in coexisting garnet (Hara et al., 2018; Spandler et al., 2003; Usui et al., 2007; Vitale Brovarone et al., 2014).

Post-peak metamorphic conditions are recorded in every lithology investigated (Figure 11a). These can be sub-divided further into a post-peak 1, and a post-peak 2 stage (Figures 10, 11a). Post-peak 1 is recorded by mica populations $\text{Ph}_{\text{MS},2a}$ and $\text{Ph}_{\text{MS},2b}$ in the mafic schists (Z16TB11 and Z16TB28), and $\text{Ph}_{\text{Cld},2}$ in the Cld-schist (Z16TB24A). The composition of these mica populations plot in a restricted P – T region at slightly lower pressure than the lawsonite-out curve (Figure 11a), at $550 \pm 15^\circ\text{C}$ and 2.00 ± 0.10 GPa with a good correspondence among phase diagram and multi-equilibrium methods (Figure 9b,e,h). Fluids are known to promote K-white mica re-equilibration (Airaghi et al., 2017; Konrad-Schmolke et al., 2011). It is therefore possible that fluid released by lawsonite breakdown favoured partial K-white mica re-equilibration by replacement reactions. The lower limit of post-peak 1 conditions is constrained at $555 \pm 15^\circ\text{C}$ and 1.55 ± 0.10 GPa, as indicated by garnet from the Cld-schist ($\text{Grt}_{\text{Cld},2}$; Z16TB24A) and the mafic schist ($\text{Grt}_{\text{MS},2}$; Z16TB11), as well as by K-white mica from the mafic schists ($\text{Ph}_{\text{MS},3a}$ in Z16TB11; $\text{Ph}_{\text{MS},3b}$ in Z16TB28). Thermodynamic results for phengite for this stage coincide when comparing phase diagram and multi-equilibrium results (Figure 9e,h), except for sample Z16TB11 for which a 1 GPa discrepancy is found (Figure 9b). Post-peak 2 conditions ($590 \pm 15^\circ\text{C}$ and 1.70 ± 0.10 GPa, Figure 11a) are constrained by garnet from Cld-schist ($\text{Grt}_{\text{Cld},3}$ in Z16TB24A), mafic schist ($\text{Grt}_{\text{MS},3}$ in Z16TB11), and mafic granofels ($\text{Grt}_{\text{MF},2b}$ in Z18TB15). The P – T conditions are similar for $\text{Grt}_{\text{MS},3}$ and $\text{Grt}_{\text{MF},2b}$ at $590 \pm 15^\circ\text{C}$ and 1.70 ± 0.10 GPa, which can indicate heating from post-peak 1 if both represent a single P – T cycle. The pressure stability condition of $\text{Grt}_{\text{Cld},3}$ is $2.05 + 0.05/-0.40$ GPa (Figure 10a,b,d), such pressure uncertainty is caused by a strong pressure dependency on minor garnet chemical variations in the model. A minimum pressure of $\sim 1.90 \pm 0.25$ GPa is constrained by the chloritoid-in curve with X_{Mg} at 0.38 (Figure 11a). The late K-white mica

TABLE 4 Lu–Hf isotope data and apparent ages

Sample fraction	Size fraction (mm)	Concentration		Isotope ratios				Final data		
		Lu (ppm)	Hf (ppm)	¹⁷⁶ Lu/ ¹⁷⁷ Hf	2SD	¹⁷⁶ Hf/ ¹⁷⁷ Hf	2SD	Lu–Hf age (Ma)	MSWD	¹⁷⁶ Hf/ ¹⁷⁷ Hf _{initial} 2SD
Grt-1	0.25–0.50	2.26	0.068	4.68700	0.0120	0.286466	0.000047			
Grt-2	>0.50	1.82	0.062	4.15700	0.0100	0.285977	0.000041			
Grt-3	0.25–0.50	2.31	0.074	4.44600	0.0110	0.286234	0.000031			
Grt-4	>0.50	1.83	0.083	3.11500	0.0080	0.284978	0.000037			
WR-1		0.336	7.26	0.006555	0.000016	0.282081	0.000015			
WR-2		0.508	6.58	0.010954	0.000027	0.282061	0.000014			
							All analyses included:	50.3 ± 0.3	1.5	0.282062 0.000010
Grt-1	>2.00	1.49	0.123	1.72300	0.0040	0.283566	0.000026			
Grt-2	>2.00	1.37	0.131	1.48600	0.0040	0.283354	0.000024			
Grt-3	>2.00	1.38	0.231	0.84580	0.0021	0.282745	0.000031			
WR-1		0.936	7.79	0.017020	0.000040	0.281988	0.000016			
							All analyses included:	49.6 ± 0.8	0.34	0.281971 0.000015
Grt-1	>0.50	2.52	0.038	9.45700	0.0240	0.290219	0.000040			
Grt-2	>0.50	3.74	0.038	14.1400	0.0400	0.294459	0.000043			
Grt-3	0.25–0.50	3.63	0.038	13.6800	0.0300	0.294014	0.000044			
Grt-4	0.25–0.50	3.62	0.034	13.7400	0.0340	0.294139	0.000050			
WR-1		0.444	7.96	0.007897	0.000020	0.281576	0.000018			
							All analyses included:	48.8 ± 0.1	1.4	0.281570 0.000018

Standard deviation (SD).

generation in the mafic schist (sample Z16TB28: $K_{\text{wm}_{\text{MS},1}}$ and $K_{\text{wm}_{\text{MS},2}}$) cannot be modelled in the phase diagram (Figure 9e) but indicates lower pressure conditions via multi-equilibrium ($P < 6$ GPa for $T < 600^\circ\text{C}$).

In the mafic schist Z16TB11, garnet porphyroblasts have comparable compositions in major elements to the rim of small garnet grains (Figure 7a,b) demonstrating that large-scale equilibrium was achieved in the intergranular medium (Lanari & Engi, 2017). This implies that the garnet porphyroblasts recorded equivalent P – T conditions than the small garnet rims and consequently grew during the post-peak 2 heating stage. This is supported by the thermodynamic modelling of garnet modes and compositions, which predicts significant new garnet growth at stage $\text{Grt}_{\text{MS},2}$ and $\text{Grt}_{\text{MS},3}$ (6 and 13 vol.%, respectively, Figure 10b). Additionally, the REE patterns of garnet porphyroblasts and rims of small grains have a comparable trend, without negative Eu anomalies (Figure 8a–c). This change in the europium budget could be due to lawsonite breakdown and epidote growth at this P – T stage (Figure 10). There is a notable difference in HREE mass fraction, with small garnet rims being more HREE enriched than the porphyroblasts. The fine interlayering of chemical domains in the mafic schist (Figure 2c) could be the origin of the variation in REE content between garnet types, which could have grown from different local bulk compositions. The rims of garnet from Cld-schist Z16TB24A are richer in MREE than the rims of fine-grained garnet in mafic schist Z16TB11 (Figure 8b,d). This difference could reflect the presence of zoisite sequestering the MREE in the schist (Konrad-Schmolke, Zack, et al., 2008; Spandler et al., 2003).

6.2 | Resorption and re-equilibration of garnet

Sharp chemical zoning is observed in the small garnets from the mafic schist (Figure 4a–d) and the Cld-schist (Figure 5) and significant variations in major elements are observed between garnet core and rim (Figure 7a). Garnet crystals with discontinuous chemical zoning comparable to those found in the TGU rocks have been interpreted as having recorded either a mono- or a poly-metamorphic history (Gaidies et al., 2006; Konrad-Schmolke, O'Brien, et al., 2008; Vance & O'Nions, 1990). In the Alps, most cases of poly-metamorphic garnet are reported from felsic metamorphic rocks, where Permian amphibolite to granulite facies garnet was partially overprinted by eclogite facies assemblages during Alpine subduction. Examples of similar garnet have been described in several localities including the Sesia Zone (Giuntoli, Lanari, Burn, et al., 2018; Giuntoli, Lanari, & Engi, 2018; Lanari et al., 2017), the Dent Blanche tectonic system (Manzotti et al., 2012), the Money unit in the Gran Paradiso Massif (Manzotti & Ballèvre, 2013), the Mt.

Emilius klippe (Angiboust et al., 2017; Hertgen et al., 2017), and in Corsica (Martin et al., 2011). A similar interpretation was given for poly-metamorphic garnet in mafic rocks in the Adula nappe, Central Alps (Herwartz et al., 2011; Sandmann et al., 2014). By contrast, mono-metamorphic garnet with complex internal textures was described in both mafic rocks in the Monviso unit (Angiboust et al., 2011, 2012, 2014; Locatelli et al., 2018; Rubatto & Angiboust, 2015), in the Tauern window, Eastern Alps (Kurz et al., 1998) and also postulated for felsic rocks in the Sesia Zone (Konrad-Schmolke et al., 2006). In the mono-metamorphic scenario, mechanisms invoked to explain the garnet textures were: (a) garnet fractional crystallization, (b) free fluid fractionation, (c) availability of water in the system, and (d) kinetics related to element supply required for garnet growth.

For the TGU samples, growth of garnet in different stages, as testified by the chemical zones, was investigated with series of simulations. For each simulation, equilibrium was assumed and pre-existing garnet was allowed to be preserved (fractionated) and/or resorbed. The water content required in the simulation to match the observed paragenesis was investigated prior to each modelling step, where equilibrium phase diagrams were calculated with pristine or fractionated bulk rock chemistry (Figures S7–S9). In every case, the best result was achieved with water saturation and was implemented as an input parameter in the simulation, which did not prevent development of successive garnet resorption and overgrowth steps matching the observations (Figure 10). Moreover, intracrystalline diffusion in garnet is sluggish below 600°C (Caddick et al., 2010) and garnet interiors are isolated from chemical equilibrium with the matrix, preventing re-equilibration (Hollister, 1966; Kohn, 2003; Spear, 1988). Consequently, interaction with an interstitial fluid, which is out of equilibrium with pre-existing garnet in a water-saturated system is the best candidate to trigger mineral reactions and mineral re-equilibration with changing P – T conditions (Ferry, 1983; Putnis, 2009). The P – T conditions of K-white mica re-equilibration by dissolution–precipitation, together with the results of thermodynamic modelling on garnet constrained fluid–rock interaction to happen during the first stage of decompression at $\sim 560 \pm 10^\circ\text{C}$ and 1.9 ± 0.1 GPa (Figure 11a).

One possibility for the release of fluids and consequent fluid–rock interaction is lawsonite dehydration, as the lawsonite out curve is crossed during the exhumation path (Figure 11a). Because lawsonite contains 11–12 wt% H_2O (Okamoto & Maruyama, 1999; Poli & Schmidt, 2002), this dehydration reaction can lead to significant water release during retrogression, as described for other metaophiolites of the western Alps (Angiboust & Agard, 2010; Angiboust et al., 2012; Groppo et al., 2009; Reinecke, 1998; Vitale Brovarone & Beyssac, 2014). In the investigated rock types, lawsonite is predicted to be stable at pressure peak conditions (Figure 10a–c) with varying modal abundance: 26 vol.% in

mafic granulites (Z16TB32), 22% in mafic schist (Z16TB11), and 1% in Cld-schist (Z16TB24A). Forward modelling using Theriak Domino simulating an isothermal decompression from the pressure peak (580°C, 2.65 GPa) to the post-peak 1 conditions (555°C, 1.55 GPa), shows variable amount of water release associated with hydrous mineral breakdown: ~0.5 wt% water release in the Cld-schist due to successive carpholite and lawsonite breakdown; ~1.5 wt% water release in the mafic schist, where Lws-out reaction produces zoisite; ~2.5 wt% water release in the mafic granulites, where lawsonite is replaced by zoisite and quartz (Figure S11). These values are above the threshold for pore connectivity (2 vol.%) that allows water migration in eclogite (Mibe et al., 2003) and will likely saturate the system in H₂O. These fluids may have circulated along preferential pathways within and across samples, such as in the garnet-rich zone in the mafic schist, where most grains have complex replacement textures.

Based on field observations and a detailed petrological study, we identified structures indicating intense deformation, such as mafic boudin stretched within the schists and veinlets with high-*P* assemblage (almandine–grossular and diopside) cross cutting the mafic granulites (Figure 2a and Figure S12a). Additionally, some schist layers contain abundant deformed clusters of garnet grains that show replacement textures, and which could be interpreted as brittle deformation of garnet in shear zones (Figure S12b). These features are comparable to what described for the high-*P* shear zones within the Monviso eclogites, and that have been linked to ductile shearing, brittle failure and fluid–rock interaction (Angiboust, Agard, et al., 2012; Locatelli et al., 2018, 2019). Also in the case of the TGU it is possible that the garnet microtextures are indicative of fluid pulses linked to seismic brecciation and intermediate-depth earthquakes.

6.3 | Garnet and the mono-metamorphic nature of the TGU

The *P*–*T* evolution of the TGU reconstructed so far could represent a single cycle of burial and exhumation. Determining this issue with certainty—and interpreting the *P*–*T* data in terms of tectonic and geodynamic processes—may prove to be difficult on the basis of *P*–*T* data alone. Garnet geochronology provides absolute age constraints needed to solve the mono- versus poly-metamorphic character of the TGU, as it has been shown in other studies in the Alps with similar complex garnet chemical zoning (Herwartz et al., 2011; Sandmann et al., 2014). In this study, the three dated samples yield similar Lu–Hf age between 50.3 ± 0.3 and 48.8 ± 0.1 Ma. In each sample the initial $^{176}\text{Hf}/^{177}\text{Hf}$ is lower than the CHUR value as expected for crustal samples. Notably, there is no resolvable age difference between sample Z16TB24A, which contains garnet with complex textures and sharp chemical

transition between core and rim, and samples where garnet shows continuous single-stage growth zoning (Z16TB28 and Z18TB15). It is unlikely that these ages are significantly influenced by diffusional re-equilibration of either Lu or Hf. Lutetium distributions show sharp and euhedral zoning and even fine oscillations, indicating primary HREE growth zoning is preserved and diffusional re-equilibration did not occur on scales larger than microns. Given that diffusivity of Hf⁴⁺ in garnet is even slower than that of Lu³⁺ (Bloch et al., 2015; Ganguly & Tirone, 1999), we interpret the Lu–Hf ages as each reflecting garnet growth during the Alpine metamorphic history of the tectonic unit. These results confirm that the lithologies of the TGU underwent a single Alpine metamorphic cycle with early garnet growth at 50–49 Ma.

Garnet porphyroblasts from the mafic schist (Z16TB11), used for GrtMod thermodynamic modelling, has the highest Lu concentration in the garnet core, whereas the rim is characterized by oscillatory and sector zoning (Figure 6b). The dated garnet porphyroblasts in the mafic schist (Z16TB28) present similar major element compositions (Figure 7b) and similar REE core to rim patterns (Figure 8a,c). In both mafic schist samples (Z16TB11 and Z16TB28), garnet porphyroblasts show HREE-depleted rims compared to their core (Figure 8a,b), possibly reflecting Rayleigh fractionation of elements that are compatible in garnet (Moore et al., 2013; Otamendi et al., 2002). We interpret the Lu distribution in Z16TB11 as representative for the dated sample Z16TB28.

As discussed above, garnet in these samples records complex zoning that represents a substantial range of *P*–*T* conditions and thus a time span. In a bulk-grain garnet analysis, and in the absence of diffusional re-equilibration of Lu and Hf, the Lu–Hf age may be biased towards zones that contribute most to the bulk-grain Lu budget (e.g. Lapen et al., 2003; Smit et al., 2010). Garnet in all samples investigated shows Lu-enriched cores and thus the Lu–Hf ages in these samples can be generally taken to represent the early stage of garnet growth, as is common in eclogite facies garnet (Lapen et al., 2003; Skora et al., 2006; Smit et al., 2010). Therefore, the Lu zoning in each garnet type (Figure 6) can further assist in assigning the age to a particular *P*–*T* stage. The three Lu–Hf ages are different outside analytical uncertainty. GrtMod thermodynamic models indicate that garnet from the mafic granulites records the early stage of TGU metamorphism. This garnet type yields the oldest age (50.3 ± 0.3 Ma), probably representing the initiation of TGU prograde metamorphism. The Lu-rich core of garnet with sharp chemical zoning from the Cld-schists is predicted stable at maximum metamorphic pressure, which consequently occurred *c.* 1.5 Ma later at 48.8 ± 0.1 Ma. Finally, major element and REE composition of garnet porphyroblasts from the mafic schist indicate growth during the re-heating stage. The garnet porphyroblasts yielded an age of 49.6 ± 0.8 Ma, which is within uncertainty

of the age of the small garnet and thus argues for rapid initial decompression and exhumation, within 1 Ma.

The ages obtained for Alpine metamorphism in this study are consistent with what was determined with various dating methods for Alpine metamorphism in the surrounding meta-ophiolites from the Western Alps: 52–40 Ma for eclogites from ZSZ on the basis of zircon and garnet dating (Amato et al., 1999; Dragovic et al., 2020; Lapen et al., 2003; Meyer et al., 2014; Rubatto et al., 1998; Skora et al., 2015), as well as Rb–Sr geochronology of mica inclusions in garnet (De Meyer et al., 2014). Comparable ages are recorded in the Monviso metaophiolite, which underwent a similar P – T evolution during Alpine subduction: 49.2 ± 1.2 Ma for garnet (Duchêne et al., 1997), and 46–45 Ma for high- P zircon rims in eclogites (Rubatto & Angiboust, 2015; Rubatto & Hermann, 2003). Garnet from eclogites from the continental slice of Etnol-Levaz yielded older ages of 61.8 ± 1.8 Ma and 52.4 ± 2.1 Ma, which could indicate a different timing of subduction-related metamorphism in this sub-units (Fassmer et al., 2016). High- P zircon in an eclogite from the Etnol-Levaz unit was dated at 47.5 ± 1.0 Ma, in agreement with other metaophiolite slices (Beltrando et al., 2010). Moreover, the 50–49 Ma Lu–Hf garnet ages obtained for all lithologies in this study are slightly younger than the Lu–Hf garnet ages described by Weber et al. (2015) from garnet in mafic granofels (56.5 ± 2.7 Ma and 58.2 ± 1.4 Ma) and favours a synchronous P – T path between TGU and ZSZ unit during Eocene.

6.4 | Geothermal gradient and comparison with surrounding tectonic units

The recognition of the P – T –time data as representing a single metamorphic cycle enables interpretation of these data in terms of tectonic and geodynamic processes. The P – T –time path of the TGU is characterized by two metamorphic maxima occurring at approximately the same temperature but at significantly different pressures (Figure 11a). Thermal structure fluctuations in subduction zones could either be linked to the thermal structure of the lithosphere itself (Peacock & Wang, 1999), to variation in the rate of subduction (Peacock & Wang, 1999), or to endothermic reactions in the subducting lithologies (Reinecke, 1998). The prograde P – T path is characterized by a steep increase in pressure of ~ 1 GPa over $\sim 90^\circ\text{C}$ (Figure 11b), indicating rapid burial with only minor thermal equilibration. The maximum P – T conditions lie on a cold apparent geotherm of $\sim 6^\circ\text{C}/\text{km}$ expected for rapid burial of cold surface sediments (Philpotts & Ague, 2009). This prograde-to-peak evolution is comparable to what was previously reported for tectonic units in the same region in the Western Alps (Figure 11b), for example, ZSZ (Angiboust et al., 2009; Bucher et al., 2005; Groppo et al., 2009; Reinecke, 1998),

Etnol-Levaz Unit (Fassmer et al., 2016), and Monviso (Angiboust, Langdon, et al., 2012). In these units, peak metamorphism is followed by isothermal decompression, for example, ZSZ in Pfulwe (Bucher et al., 2005), Lago di Cignana (Reinecke, 1998), and Täschalp (Barnicoat & Fry, 1986) and for some continental outliers (Angiboust et al., 2017; Fassmer et al., 2016). Such P – T trajectory is possible if the tectonic units follow a rapid exhumation after decoupling from the descending slab (Philpotts & Ague, 2009; Rubatto & Hermann, 2001). Numerical simulations that report pressure and temperature pathways of rocks during collision and subsequent exhumation correlate an isothermal pressure drop with the onset of slab break off (Vogt & Gerya, 2014), which may be the geodynamic process governing initial exhumation of the TGU as well.

After a first stage of near isothermal decompression (post-peak 1), the exhumation path of the TGU reflects a minor, almost isobaric, heating episode lying on a slightly hotter geotherm of $\sim 10^\circ\text{C}/\text{km}$ (post-peak 2). The apparent pressure increase between post-peak 1 and post-peak 2 is not considered significant, since it is only constrained by the chloritoid isopleth, which could also reflect minor inconsistencies in the thermodynamic data. The β -shaped P – T path for TGU is consistent with the exhumation paths proposed for the ZSZ (Figure 11b; e.g. Angiboust et al., 2009; Bucher et al., 2005). Similar β -shaped P – T paths have been reported for many orogens indicating that thermal pulses during exhumation are potentially a common feature. Examples of β -shaped P – T paths are reported for many HP terranes worldwide (Faryad et al., 2019; Gao et al., 2011; O'Brien, 2000; Whitney et al., 2011; Wilke et al., 2010), including the Tauern window in the Eastern Alps (e.g. Kurz et al., 2008; Ratschbacher et al., 2004), the Lepontine Dome in the Central Alps (Wiederkehr et al., 2008), the Internal Penninic Nappes in the Western Alps (e.g. Borghi et al., 1996; Brouwer et al., 2002; Rubatto & Hermann, 2001), and the Upper Penninic Nappes in the Western Alps (Beltrando et al., 2007; Messiga & Scambelluri, 1991). Various mechanisms have been invoked to affect geothermal gradient and explain the subsequent reheating step, such as radiogenic heat production of additionally accreted continental blocks (Wiederkehr et al., 2008), slab breakoff (Brouwer et al., 2002; Kurz et al., 1998), continental collision after the end of oceanic lithosphere subduction (Borghi et al., 1996), mantle delamination (Henk et al., 2000), slowing exhumation rate (e.g. Wilke et al., 2010), asthenospheric inflow ahead of the spreading ridge (Abers et al., 2016), transient storage at MOHO depth during exhumation (Walsh & Hacker, 2004) and multiple, short-lived burial-exhumation cycles related to orogen-scale switches between shortening and extensional deformation (Beltrando et al., 2007). Recent numerical modelling of exhumation mechanisms has identified advective and conductive heating from the mantle, rather than radiogenic or shear heating, as

best candidates to explain rapid (short-lived) heating of rocks during exhumation (Sizova et al., 2019). The numerical results show β -shaped P - T -time paths with re-heating of up to 200°C at various pressure (0.5–1.5 GPa) and involve asthenospheric upwelling related to slab rollback or slab bending, or crustal exhumation paired with slab breakoff, as prime driver for late orogenic heating during exhumation (Sizova et al., 2019). In the case of the TGU, the imprint of such processes would be relatively strong, as this unit is structurally located on top of the subducted oceanic crust and is therefore located closest to the slab-mantle interface where thermal anomalies would be most intense.

6.5 | Comparison with previous findings

Complex garnet textures with sharp chemical zoning in the TGU were previously interpreted as the result of polycyclic metamorphism (Bucher et al., 2019). The garnet zoning of the schist samples investigated in this study (Z16TB11 and Z16TB24A) shows good correspondence with those observed by Bucher et al. (2019). In both cases, a marked compositional difference is present between garnet core and rim (Figure 7a). However, in this study Lu–Hf geochronological data obtained for garnet from different lithologies returned coherent Alpine ages in every sample. Particularly, the Alpine age of Cld-schist (Z16TB24A), where garnet has a discontinuous core–rim zoning, excludes a pre-Alpine (Permian) metamorphic stage as suggested by Bucher et al. (2019). This study demonstrates that such textural record is compatible with a mono-metamorphic evolution of the TGU. As a main consequence, there is no longer discrepancy for the TGU Alpine peak P - T conditions between the mafic granulites at 2.2 ± 0.1 GPa and $580 \pm 50^\circ\text{C}$ (Weber & Bucher, 2015), and the schist at ~ 1.7 GPa and $\sim 530^\circ\text{C}$ (Bucher et al., 2019). The garnet core in the schists (granulitic according to Bucher et al., 2019) is re-interpreted as representing Alpine peak- P conditions, corresponding to the maximum P - T conditions inferred in the TGU mafic granulites at 2.65 ± 0.10 GPa and $580 \pm 15^\circ\text{C}$ (this study; Weber & Bucher, 2015). Likewise, the maximum P - T conditions for the metapelites in Bucher et al. (2019) corresponds to the initiation of the re-heating stage after the first step of exhumation, rather than the Alpine peak pressure for these rocks.

Compositional maps of the garnet in the Cld-schist (Z16TB24A) (Figure 4a–d) and the matrix garnet of the mafic schist (Z16TB11) (Figure 5) show a sharp chemical boundary in major elements between core and rim at the high lateral resolution of the EMPA map (~ 1 μm in garnet, see Lanari & Piccoli, 2020). Pristine chemical zoning could be modified by diffusional relaxation, which is dependent on cooling rate and grain size (Dodson, 1973), because there is a strong dependence between cation diffusivity and temperature in

minerals (Caddick et al., 2010). The sharp zoning shows no evidence of diffusional re-equilibration of relatively fast diffusing divalent cations at sharp chemical interfaces in garnet, such as Fe, Mg, and Mn (Carlson, 2006; Li et al., 2018; Perchuk et al., 2009). This result is inconsistent with the interpretation that the TGU resided at 530 – 580°C for *c.* 10 Ma (Bucher et al., 2019), but rather argue for a fast P - T cycle as suggested by our Lu–Hf data.

7 | CONCLUSIONS

The P - T path reconstructed for the TGU unit combined with Lu–Hf dating of garnet reveals a coherent mono-metamorphic history during Eocene, which is in agreement with the surrounding SZS metaophiolites. After a first stage of decompression, the exhumation path of the TGU records a reheating event. Complex garnet textures are thus reconciled with a mono-metamorphic evolution and a β -shaped P - T path. The thermal pulse is best explained by upwelling of hot asthenospheric mantle, favoured by the structural position of the TGU located atop of the subducted oceanic crust and therefore closer to the slab-mantle interface. During exhumation of the TGU, lawsonite dehydration occurring in both the mafic schist and the mafic granulites is a main fluid source for fluid–rock interaction at high pressure. The presence of significant amounts of fluids leads to complex texture development in garnet from the schist, K-white mica re-equilibration in the schists and vein formation in the mafic granulites.

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SUPPORTING INFORMATION

Additional supporting information may be found online in the Supporting Information section.

Figure S1. Quantified LA-ICP-MS compositional maps in garnet, all maps show $\mu\text{g/g}$ abundance.

Figure S2. Quantified LA-ICP-MS compositional maps in

garnet, all maps show $\mu\text{g/g}$ abundance.

Figure S3. Quantified LA-ICP-MS compositional maps in garnet, all maps show $\mu\text{g/g}$ abundance.

Figure S4. BSE image of garnet.

Figure S5. EPMA profile across chloritoid grains.

Figure S6. Ternary chemical diagram showing the range of garnet composition in $X_{\text{grs}}-X_{\text{alm}}-X_{\text{prp}}$.

Figure S7. Pseudosection of sample Z16TB11 with garnet $\text{Grt}_{\text{MS},1}$ P - T stable condition.

Figure S8. Pseudosection of sample Z16TB11 with garnet $\text{Grt}_{\text{MS},2}$ P - T stable condition.

Figure S9. Pseudosection of sample Z16TB11 with garnet $\text{Grt}_{\text{MS},3}$ P - T stable condition.

Figure S10. Pictures of garnet separates for Lu-Hf dating.

Figure S11. P - T diagrams presenting variation of H_2O in solids with superimposed pseudosections.

Figure S12. Outcrop pictures and scanned thin section images illustrating fluid-rock interaction in TGU.

Table S1. Garnet major element average composition.

Table S2. LA-ICP-MS trace element of garnet.

Table S3. K-white mica major element average composition.

Table S4. Chloritoid major element chemical composition.

Table S5. LA-ICP-MS trace element of rutile.

Table S6. Reactive bulk rock chemistry evolution using GrtMode.

Table S7. LA-ICP-MS major and trace element bulk rock analyses.

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