#### **ORIGINAL PAPER**



# Monazite in the eclogite and blueschist of the Svalbard Caledonides: its origin and forming-reactions

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# Abstract

High-pressure low-temperature rocks from Svalbard are an excellent target for studying metamorphic reactions in Phanerozoic subduction zones. This study reveals the presence of monazite in an eclogite and a blueschist from the Vestgötabreen Complex, southwestern Svalbard. In order to investigate the monazite-forming reaction, we obtained pressure-temperature estimates coupled with U-Pb and Lu-Hf dating. Combined geothermobarometry allows to constrain three evolutionary stages of garnet growth in the eclogite: nucleation  $(1.6 \pm 0.3 \text{ GPa at } 460 \pm 60 \text{ °C})$ , peak-pressure  $(2.3 \pm 0.3 \text{ GPa at } 507 \pm 60 \text{ °C})$ , and peak-temperature (2.1  $\pm$  0.3 GPa at 553  $\pm$  60 °C). A zircon age of 482  $\pm$  10 Ma is interpreted to belong to the prograde part of the pressure-temperature path. Monazite forms inclusions within garnet rims, or it is surrounded by allanite and apatite, altogether forming pseudomorphs of a tabular shape in the matrix. Textures, geothermobarometry and geochronology support the conclusion the monazite formed under high-pressure conditions at  $471 \pm 6$  Ma. We propose that the monazite crystallization in the eclogite happened due to a decomposition of accessory phases during the decompression after peakpressure of the metamorphic cycle. Monazite in the blueschist occurs as inclusions in garnet cores and gives an indicative age of  $486 \pm 6$  Ma, which is interpreted to reflect the prograde growth of the garnet. Lu–Hf garnet dating resolves an age of peak-pressure metamorphism in the blueschist at  $471.1 \pm 4$  Ma under conditions of  $2.0 \pm 0.03$  GPa and  $500 \pm 30$  °C. The Vestgötabreen Complex provides evidence for an early Ordovician modern-style subduction system in the proximity of the Baltica margin. Hence, this study also supports the tectonic models that favour a mixed Baltican and Laurentian provenance of south-western Svalbard.

Keywords Eclogite  $\cdot$  Blueschist  $\cdot$  Monazite  $\cdot$  High-pressure low-temperature metamorphism  $\cdot$  Vestgötabreen Complex  $\cdot$  Svalbard

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# Introduction

Monazite is a common accessory mineral in metaigneous peraluminous felsic rocks and metapelites that formed under high-pressure (HP) conditions (e.g., Spear and Pyle 2002; Finger and Krenn 2007; Williams et al. 2007; Krenn et al. 2009; Stepanov et al. 2014; Schulz 2021). However, it is very scarce in metamafic rocks such as eclogites sensu stricto (e.g., Schulze et al. 2015). This phenomenon of the presence of monazite applies to felsic and mafic rocks of all metamorphic grades and results mostly from the bulk content of CaO (e.g. Spear 2010). In general, the lower the Ca activity is in the bulk rock, the more likely it is that monazite will form as the light rare earth element (LREE) phase. This simple relationship has been documented in several natural examples where monazite-allanite relationships have been described as a function of Ca-bearing fluid activity (e.g. Finger et al. 1998; Broska and Siman 1998; Gasser et al. 2012; Kośmińska et al. 2020). Additional important constrains for the formation of monazite are the temperature gradient along which a rock evolves, as well as the pH of the metasomatizing fluid, as demonstrated in experimental works (e.g. Budzyń et al. 2017). Taking all these factors into account, it is clear that mafic rocks would favor epidote supergroup minerals as the main LREE carriers, while peraluminous granitoids and metapelites would favor monazite with the absolute pressure of the metamorphic process playing a less important role. However, in some rare cases, monazite can occur in metamafic rocks (e.g. Schulze et al. 2015). To better understand the enigmatic occurrence of monazite in eclogite and blueschist and its relationship to other major and accessory phases, we investigated monazite-bearing high-pressure low-temperature (HP-LT) rocks from the Vestgötabreen Complex on Svalbard in terms of their pressure-temperature (P-T) evolution, timing of metamorphism and potential fluid infiltration.

The main LREE carriers in mafic HP lithologies are allanite, epidote and to lesser degree lawsonite (e.g., Hermann 2002; Poli and Schmidt 2004; Gieré and Sorensen 2004; Spandler et al. 2003; Martin et al. 2014; Whitney et al. 2020). Lawsonite is a mineral typical of HP-LT lithologies but is rarely preserved during transport back to the earth surface (Clarke et al. 2006; Tsujimori and Ernst 2014; Hamelin et al. 2018; Chapman and Clarke 2021). Several reactions have been proposed for lawsonite breakdown which, together with reaction textures and trace element analyses, help define the presence of lawsonite during prograde evolution. A less common LREE-bearing mineral known from HP-LT rocks is florencite (CeAl<sub>3</sub>(PO<sub>4</sub>)<sub>2</sub>(OH)<sub>6</sub>; Chopin and Monie 1984; Janots et al. 2006; Gautiez-Putallaz et al. 2016; Hoschek 2016). Florencite has been described from HP units in the Alps (e.g., Gran Paradiso—Chopin 1979; Gabudianu Radulescu et al. 2009, Monte Rosa—Chopin and Monié 1984) and ultra-HP units (Dora-Maira—Gautiez-Putallaz et al. 2016), blueschist-facies metapelites from Morocco (Janots et al. 2006), and HP-LT phyllites from Svalbard (Barnes et al. 2020a). Experimental data suggests that florencite is stable up to 2.5 GPa at 500 °C and 1.5 GPa at 700 °C (Barthélemy and Chopin 2006).

The blueschist- and eclogite-facies rocks of the Vestgötabreen Complex represent an early Paleozoic mélange (Ohta et al. 1986; Hirajima et al. 1988; Dallmeyer et al. 1990; Bernard-Griffiths et al. 1993; Labrousse et al. 2008) formed along a low geothermal gradient (Agard et al. 2005). HP-LT assemblages are well preserved in the units with almost no signs of retrogression. Although the complex has been recognized for decades, detailed petrochronological studies have not been performed. The pressure–temperature–time (P–T–t) history of eclogites and blueschists from the Vestgötabreen Complex is constrained in this study using a combination of geochronological and petrological methods. We describe monazite from HP-LT eclogite and propose that monazite crystallization potentially occurs as a result of florencite and lawsonite decomposition.

# **Geological setting**

# **Arctic and Svalbard**

Svalbard is an archipelago in the high Arctic, located between 74° and 81° N (Fig. 1a). Svalbard, along with Pearya (northern Ellesmere Island) and Arctic Alaska terranes, was part of a Paleozoic terrane which formed on the northern Laurentian margin (see more details in McClelland et al. 2022). The pre-Devonian rocks of Svalbard are traditionally divided into three basement provinces, namely the Eastern, Northwestern, and Southwestern provinces (e.g., Harland 1997; Gee and Teben'kov 2004; Dallmann 2015; Fig. 1b), based on their tectonostratigraphy and age. HP rocks occur in the Northwestern province (Richarddalen Complex, Gromet and Gee 1998; Elvevold et al. 2014) and are spread along the western margin of the Southwestern province (Labrousse et al. 2008; Kośmińska et al. 2014). Here, we focus on blueschist and eclogite from the Vestgötabreen Complex of the Southwestern province which crops out on the Oscar II Land. The Southwestern province is composed of Mesoproterozoic to Neoproterozoic rocks (Dallmann 2015; Ziemniak et al. 2019; Wala et al. 2021). The rocks locally experienced several episodes of metamorphism and/or deformation: during the ca. 640 Ma Torellian event (Majka et al. 2008, 2014), in the Ordovician during the Caledonian orogeny (Faehnrich et al. 2020; Barnes et al.



Fig. 1 a Location of the study area; b simplified map of Svalbard (Gee and Tebenkov 2004; Dallmann 2015). N Nordenskiöld Land, O Oscar II Land, S Sarsøyra area; c geological map (Bergh et al. 2003)

2020a) and during the late Devonian—early Carboniferous Ellesmerian orogeny (e.g., Kośmińska et al. 2020). Recent detrital zircon studies suggest that the Southwestern Svalbard province is a complex terrane with both Laurentian and

Baltican affinities (Gasser and Andresen 2013; Ziemniak

et al. 2019; Olierook et al. 2020; Wala et al. 2021).

#### **Oscar II Land**

Oscar II Land (Fig. 1b, c) consists mainly of Meso- and Neoproterozoic rock units including carbonates, phyllites and quartzites (e.g. Dallmann 2015; Majka and Kośmińska 2017) of the Daudmannsodden and St. Jonsfjorden groups (Hjelle et al. 1999) as well as Cryogenian tilloids and platform-type carbonates of the Comfortlessbreen Group. Greenstone and greenschist locally form layers within low grade Meso-Neoproterozoic metasedimentary rocks that are overthrust by the HP-LT Vestgötabreen Complex in the Motalafjella area (e.g. Horsfield 1972; Ohta 1979; Labrousse et al. 2008; Fig. 1c). The Vestgötabreen Complex is unconformably overlain by the Late Ordovician to Early Silurian Bullbreen Group, which consists of conglomerate,

of the Vestgötabreen Complex key locality-Motalafjella area, sample locations are marked with stars

carbonate, sandstone and shale (Scrutton et al. 1976; Armstrong et al. 1986; Fig. 1c). The Bullbreen Group was only slightly metamorphosed (~300 °C based on color index in conodonts) and folded during pre-Devonian time. The exact age of this deformation is not constrained (Dallmeyer et al. 1990; Ohta 1994), but Barnes et al. (2020a) suggest that the thermal overprinting of the Bullbreen Group is related to thrusting of HP-LT rocks at ca. 430–400 Ma.

Blueschists and eclogites of the Vestgötabreen Complex (Ohta et al. 1986; Kanat and Morris 1988) are the best-preserved Paleozoic HP-LT lithologies on Svalbard and in the Caledonides in general, and are one of the most northern exposures of this type of rocks in the world. The Vestgötabreen Complex type locality is in the Motalafjella and Skipperryggen nunataks (Fig. 1c). Serpentinites in the Sarsøyra area of Oscar II Land (Ohta et al. 1995) and retrograde blueschists in Nordenskiöld Land (Kośmińska et al. 2014) may represent northern and southern continuations, extending the HP-LT belt to ~ 130 km in length. The Vestgötabreen Complex is divided into two structural units, the Upper unit and the Lower unit. The Upper unit is represented by garnet-bearing schist and carbonate with lenses of blueschist and eclogite (Fig. 1c). The Lower unit consists of HP-LT metasedimentary rocks, phyllites, calc-schists and magnesites with lenses of serpentinite, greenstone and metacarbonate (Fig. 1c). Typical HP-LT assemblages containing lawsonite, pumpellyite and Fe-Mg carpholite have been described in the Lower unit metasedimentary rocks and phyllites (Hirajima et al. 1984, 1988; Agard et al. 2005). The Lower and Upper units yield different P-T conditions. However, both units plot along a similar low geothermal gradient. Eclogites from the Upper unit yield peak pressure conditions of 1.8-2.4 GPa at 580-640 °C (Hirajima et al. 1984, 1988), whereas P-T estimates for phyllites from the Lower unit are up to ~ 1.6 GPa at 380-450 °C (Agard et al. 2005). These conditions correspond to a subduction geotherm of approximately 7-8 °C/km, providing evidence for a cold subduction regime in the Paleozoic (Agard et al. 2005). The timing of HP metamorphism is accepted as Early Ordovician (around 470 Ma; Horsfield 1972; Dallmeyer et al. 1990; Bernard-Griffiths et al. 1993), but modern geochronological context-sensitive in-situ studies have not been performed on HP lithologies of the Upper unit. The first isotopic dating attempt of the Vestgötabreen Complex was performed by Horsfield (1972) using K-Ar dating of white mica from various Upper unit lithologies including garnet-glaucophane-muscovite schist, muscovite-epidote-garnet-pyroxene schist, and a quartz-muscovite-garnet schist. This study yielded dates ranging from  $475 \pm 14$  to  $402 \pm 14$  Ma. The younger dates were interpreted to record a late Caledonian overprint, whereas the older dates were thought to be white mica crystallization ages. Dallmeyer et al. (1990) dated white mica from HP schists of the Upper unit using the <sup>40</sup>Ar/<sup>39</sup>Ar technique. Their study provided plateau ages ranging from  $475 \pm 2$ to  $470 \pm 2$  Ma for phengitic mica and from  $461 \pm 2$  to  $452 \pm 2$  Ma for paragonitic mica. Bernard-Griffiths et al. (1993) performed pioneering dating of eclogite using the U-Pb system in zircon. Regression of four multi-grain size fractions defined a lower intercept of 476 + 30/-29 Ma (MSWD = 2.4) which was interpreted to represent a Pbloss event related to HP metamorphism. Recently, Barnes et al. (2020a) published in-situ <sup>40</sup>Ar/<sup>39</sup>Ar dating results of white mica from both the Upper and Lower units using different white-mica-bearing schists, blueschists and phyllites. They obtained three age populations for the Upper unit:  $493 \pm 7$  Ma,  $476 \pm 2$  Ma,  $440 \pm 6$  Ma, and four age populations for the Lower unit:  $454 \pm 6$  Ma,  $426 \pm 4$  Ma,  $398 \pm 6$  Ma,  $365 \pm 11$  Ma. They interpreted the  $476 \pm 2$  Ma age to represent cooling during exhumation after peak P. which was followed by juxtaposition of the Upper Lower units at  $454 \pm 6$  Ma. The younger populations are interpreted as a late tectonothermal event potentially related to the Ellesmerian Orogeny. Finally, in-situ Th-U-total Pb

monazite dating of phyllites from the Lower unit also suggests thermal overprinting during the late Silurian–Devonian (Barnes et al. 2020a).

# Methods

#### **Electron microprobe**

Wavelength-dispersive spectroscopy (WDS) analyses of mineral chemistry as well as compositional maps of garnet, amphibole and monazite were performed using two electron microprobes (EMP): JEOL JXA8530F Hyperprobe at the Uppsala University, Sweden and JEOL JXA-8230 Superprobe at AGH University of Science and Technology, Kraków, Poland. The operating conditions for quantitative analyses were as follows: a 10-20 nA beam current with 15 kV accelerating voltage, counting times of 10 s on peaks and 5 s on background positions and focused or defocused to 5 µm beam depending on the mineral. The JEOL JXA-8230 Superprobe at AGH was used to collect the WDS compositional maps under the following analytical conditions: a 100 nA beam current with 25 kV accelerating voltage, a dwell time of 100 ms, and 0.1-0.3 µm step size for monazite maps; a 100 nA beam current with 15 kV accelerating voltage, a dwell time of 50 ms, and 0.5 µm step size for garnet and amphibole maps. Both synthetic and natural standards were used for calibration. BSE images and X-ray maps of monazite were collected using the Jeol JXA-8230 Superprobe at AGH University of Science and Technology, Kraków, Poland. BSE images of zircon were collected using the same instrument, and CL images were obtained using a scanning electron microscope (SEM), FEI QUANTA FEG 200, at AGH. Additional, major-element distribution maps of garnet were carried out using a JEOL 8200 Superprobe at the Institute of Geosciences in Bonn. They were obtained with a beam current of 50 nA, an acceleration voltage of 15 kV and a measurement time of 100 ms per point. Mineral abbreviations are after Whitney and Evans (2010).

The Cameca SX-100 electron microprobe at the Rensselaer Polytechnic Institute (Troy, NY, USA) was used to measure Zr concentrations in rutile. The operating conditions were as follows: 200 nA beam current, 15 kV accelerating voltage, counting times of 300 s on peaks and 150 s on  $\pm$  background positions. Synthetic zircon and rutile were used as standards.

#### Raman spectroscopy and elastic thermobarometry

Quartz inclusions in garnet were measured in  $\sim 80 \,\mu\text{m}$  thick sections. The measured inclusions were crack-free, close to spherical shape, and > 3 radii away from other inclusions. The Raman spectra were collected at Syracuse University

(Syracuse, NY, USA) using a Renishaw inVia Raman microprobe spectrometer system at room conditions of 1 bar and 25 °C. A 532 nm laser was focused onto the centers of inclusions using a  $100 \times \text{objective}$  (N.A. = 0.9). We operated the instrument in high confocality mode to increase the signal from quartz inclusions. Raman shifted light was backscattered (180° geometry) and statically dispersed using 1800 groove/mm grating onto a charged-couple detector. Spectra were acquired for 20 s. The Raman instrument was calibrated against the Rayleigh scattered light from the 532 nm laser, Ne lines and the 520.5 cm<sup>-1</sup> Raman band of a silicon standard. A synthetic quartz crystal cut perpendicular to the c axis was measured as a reference material throughout analytical sessions to determine the exact positions of the three largest peaks located at approximately 128 cm<sup>-1</sup>,  $206 \text{ cm}^{-1}$ , and  $465 \text{ cm}^{-1}$  in the quartz Raman spectra. The quartz inclusions are strained, which causes changes to the peak positions. We determined  $\Delta \omega$  (e.g.,  $\Delta 465 \text{ cm}^{-1}$ ) by subtracting the positions of the three main peaks of the reference material from the peak positions of the inclusions. The  $\Delta 128 \text{ cm}^{-1}$ ,  $\Delta 206 \text{ cm}^{-1}$ , and  $\Delta 465 \text{ cm}^{-1}$  values were used in stRAinMAN software (Angel et al. 2019) to calculate the amount of strain in the quartz inclusions. We used the equations of state for quartz (Angel et al. 2017 curved boundary model) and almandine (Angel et al. 2022) and the entraPT software (Mazzucchelli et al. 2021) to convert strains to Pincl and plot isomekes.

#### In-situ zircon, monazite and allanite dating

Zircon, monazite and allanite U-Th-Pb and trace element analyses were conducted in-situ by secondary ion mass spectrometry (SIMS) using the sensitive high-resolution ion microprobe with reverse-geometry (SHRIMP-RG) at the Stanford-US Geological Survey Micro-Analysis Center (SUMAC), Stanford, California. The analytical data are presented in Tables S1-S5 (Kośmińska et al. 2023) and backscattered electron (BSE) images of the dated zircon grains are presented in the Supplemental materials S1. All analyses were performed in-situ on pieces of polished thin sections that were trimmed with a diamond wire saw and embedded in epoxy. Zircon grains were characterized in the thin sections prior to mounting using transmitted and reflected light microscopy, backscattered electron (BSE) and cathodoluminescence (CL) imaging. Similarly, monazite and allanite were imaged using light microscopy, BSE photos and X-ray compositional maps. Reference materials co-mounted with the samples for U/Pb fractionation corrections and concentration include Temora-2 zircon (418.37  $\pm$  0.14 Ma; Mattinson 2010) and MAD-559 (3940 ppm U; Coble et al. 2018), 44,069 monazite (424.9±0.4 Ma; 2500 ppm U; Aleinikoff et al. 2006) and NAM (1274 ppm U; Aleinikoff et al. 2012) and Arandal allanite (1060 Ma; 431 ppm U based on an in-house intercomparison with TARA allanite 415 Ma; Gregory et al. 2007). The analytical procedure followed Barth and Wooden (2006, 2010). Secondary ions were generated using an  $O^{2-}$  primary ion beam accelerated at 10 kV with intensities of 0.7–1.2 nA for U/Pb analysis of zircon, 7.3–10.0 nA for U/Pb analysis of monazite and allanite, and 0.6–1.0 nA for trace element analysis of all three phases. Spot size was ~ 18 × 20 µm in diameter and the ablation ~ 2 to 3 µm deep. Trace element analyses were performed at the same site after U/Pb analysis with a mass resolution of 6800–7200 (10% peak height).

Data reduction follows Ireland and Williams (2003) using the MS Excel add-in program Squid 2.51 (Ludwig 2009). The error for the  ${}^{206}$ Pb/ ${}^{238}$ U calibrations for zircon, monazite and allanite based on repeated measurements of reference materials were 1.0%, 1.9% and 1.0% ( $2\sigma$ ), respectively. Initial common Pb isotopic compositions were approximated from Stacey and Kramers (1975) for zircon and monazite. Allanite initial common Pb is assigned based on data defined 3-dimensional regression of the data. No additional error was propagated for the uncertainty in the common Pb composition. <sup>206</sup>Pb/<sup>238</sup>U ages for zircon and monazite were corrected for common Pb using the measured <sup>207</sup>Pb/<sup>206</sup>Pb (see Ireland and Williams 2003). All reported <sup>206</sup>Pb/<sup>238</sup>U and  $^{207}$ Pb/ $^{206}$ Pb model ages and uncertainties (2 $\sigma$ ) include the error summed in quadrature from the external reproducibility  $(1\sigma SD)$  of the reference materials during the individual analytical session (14-24 h). Plots and three-dimensional regression of allanite data to determine total U isochron ages utilized the MS Excel add-in program Isoplot 3.75 (Ludwig 2012). Although the age for Arandal has not been precisely measured independently (e.g., ID-TIMS), this uncertainty is minimal compared to the relatively large uncertainty in the unknown allanites in this study due to the high common Pb concentration. Zircon trace element concentrations were calibrated against the zircon reference material MAD-559 (Coble et al. 2018). Monazite and allanite trace element concentrations were calculated relative to standard 44,069 and NAM (Aleinikoff et al. 2012). The estimated errors ( $1\sigma$ standard deviation) are based on repeated analysis of reference materials 3-10% for P, Y, Hf, Th, and U and the REEs, except for La (20%).

#### Laser ablation analyses of garnet

Trace element profiles through selected garnet grains were conducted with a Thermo Scientific XSeries 2 Q-ICP-MS, connected with a Resonetics RESOlution M50-E 193 nm Excimer Laser Ablation System at the Institute of Geosciences, Bonn. Spot size was 44  $\mu$ m for sample KK14-55 and 58  $\mu$ m for sample KK12-12E. Laser repetition rate was 10 Hz and laser fluency at the sample surface was 7 J/ cm<sup>2</sup> in measurements on both samples. Spots were chosen

manually to avoid inclusions. <sup>29</sup>Si, <sup>43</sup>Ca, <sup>47</sup>Ti, <sup>55</sup>Mn, <sup>91</sup>Zr, <sup>177</sup>Hf, <sup>178</sup>Hf, <sup>180</sup>Hf, and REE were analyzed, measuring for 30 s on the gas background and for 30 s on each measuring point. <sup>29</sup>Si was used as an internal standard and NIST612 as an external standard for count rate normalization (Jochum et al. 2011). Normalized count rates were converted into concentrations using the procedure after Longerich et al. (1996).

#### Lu-Hf isotope measurements

The method used for preparation and measurement of Lu and Hf in mineral separates and whole rock splits is described in detail in Fassmer et al. (2017), Miladinova et al. (2018), and Hauke et al. (2019). Whole rock splits were digested with the tabletop procedure after Münker et al. (2001) and Lagos et al. (2007), and also in a Parr bomb. Sample preparation was done at the Institute of Geosciences in Bonn and sample digestion, column chemistry and isotope measurements were conducted at the Institute for Geology and Mineralogy in Cologne. A Thermo Scientific Neptune Multicollector ICP-MS was used for isotope measurements.

# Results

Field examination of the Vestgötabreen Complex was carried out on the Skipperryggen nunatak and on the NE side of the Motalafjella ridge (Figs. 1c, 2). Eclogite and blueschist form boudins and layers included within garnet-bearing white mica schist (Fig. 2). Several samples of garnet-bearing HP lithologies were collected from the Upper unit of the Vestgötabreen Complex. Based on microscopic observations of more than 20 thin sections, two of the freshest and most representative samples of eclogite and blueschist were chosen for this study (GPS coordinates of eclogite—KK14-55 78° 26' 28.0" N 12° 52'

55.6" E an blueschist—KK12-12E 78° 27′ 34.4" N 12° 48′ 19.0" E, Fig. 1c).

# **Mineral chemistry**

Eclogite consists of subhedral garnet (20 vol %) scattered in a coarse-grained, very weakly foliated matrix composed mainly of omphacite (70 vol %) with minor sodic amphibole (5%), white mica (3%), epidote (2%), rutile and quartz (Figs. 3a, 4). The HP assemblage is well preserved with little evidence for retrogression (chloritization of garnet). Garnet porphyroblasts (up to 2 mm) contain inclusion-rich cores with inclusions of omphacite, glaucophane, rutile, quartz, white mica, apatite and epidote (Fig. 4a). Secondary chlorite is partly replacing garnet (Fig. 4a, f, h). Two distinctive garnet generations were recognized based on the chemical X-ray maps. Garnet-I forms cores of the porphyroblasts. Its composition is characterized by the range of Alm<sub>65-75</sub>Prp<sub>13-21</sub>Grs<sub>5-16</sub>Sps<sub>0-2</sub> (Fig. 5a, b; Table 1). Garnet-I displays a typical prograde zonation with decreasing  $X_{Sns}$  and  $X_{Alm}$  towards the rim (from 2 to 0 mol% and from 75 to 65 mol%, respectively) and relatively flat, low  $X_{Grs}$ content (~5 mol %). Spessartine content is low and shows a bell-shaped profile (Fig. 5b). X<sub>Prn</sub> slightly decreases in the rim direction. Garnet-II forms an outer rim across a very sharp contact with garnet-I that is clearly visible on the Ca map (Fig. 5a). The outer rims are characterized by  $Alm_{56}Prp_{26}Grs_{17}Sps_1$ , show oscillatory zonation of  $X_{Sps}$ , and display significantly higher X<sub>Grs</sub> (up to 20 mol%) content than garnet-I. Clinopyroxene is classified as omphacite. Omphacite-I forms inclusions in garnet-I and is characterized by jadeite varying from 38 up to 53 (Table 1, Fig. 5d-f). Omphacite-II occurs as rare inclusions in garnet-II and in the matrix with a jadeite content ranging from 31 to 49. Omphacite in the matrix shows oscillatory zonation (Fig. 4d, e). Sodic amphibole is represented by glaucophane that occurs as inclusions in garnet porphyroblasts or forms euhedral crystals in the omphacitic matrix (Fig. 4a-c). It shows compositional chemical zonation with cores characterized by  $X_{Fe2+} = 0.24-0.35 (X_{Fe2+} = Fe^{2+}/Fe^{2+} + Mg)$  and rims with



Fig. 2 a Garnet-white-mica bearing schist which is the dominant lithology hosting blueschists and eclogites; b lens of blueschist surrounded by HP schist; c eclogite

Fig. 3 Thin sections scans and petrogenetic table showing the main mineral assemblages for eclogite (a) and blueschist (b). Three metamorphic stages are marked: prograde, peak and retrograde



 $X_{Fe2+} = 0.38-0.48$  (Fig. 5g, Table 1). Phengite is scarce and mostly forms flakes enclosed in garnet and in the omphacitic matrix (Fig. 4f). Phengite inclusions within garnet-I are characterized by silica content varying from 3.38 to 3.45 apfu (atom per formula unit, Table 1), whereas phengite included in garnet-II shows silica content from 3.43 to 3.47 apfu. The silica content ranges from 3.40 to 3.43 apfu for matrix phengite. Phengite is generally characterized by low Na contents of 0.02-0.04 apfu. Epidote mostly forms inclusions in garnet-I and shows X<sub>Fe3+</sub> values of 0.24-0.29  $(X_{Fe3+} = Fe^{3+}/Fe^{3+} + Al; Fig. 4a, g)$ . Rutile is present in the matrix and as inclusions in garnet (Fig. 4a, c). Quartz forms tiny (< 30 µm) inclusions mostly within the garnet-I. Chlorite occurs as inclusions within garnet-I or fills cracks within garnet porphyroblats. Apatite occurs as inclusions within garnet porphyroblasts and in the matrix. The eclogite is cut by abundant late veins composed mainly of carbonates, chlorite and feldspar (Fig. 4i). Lawsonite was not recognized in the studied sample but has been described from the cores of garnet in similar eclogites studied by Hirajima et al. (1988).

Blueschist is composed of glaucophane (50%), garnet (25%), white mica (15%), epidote (7%), chloritoid (3%), and minor amounts of rutile, quartz, hematite, apatite, and carbonate (Figs. 3b, 6). Garnet forms subhedral

porphyroblasts (up to 2 mm) containing epidote, chloritoid, phengite, paragonite, rutile, and hematite inclusions which occur mostly within the cores (Fig. 6a). X-ray chemical maps of garnet from a blueschist revealed prograde chemical zoning defined by decreasing Mn, Fe and Ca and increasing Mg toward the rim (Fig. 7a, b). Grossular and almandine show only minor decreases in the inner part of the grain, and sharply decrease in the outer part. Garnet composition ranges from Alm<sub>60-66</sub>Prp<sub>9-20</sub>Grs<sub>19-24</sub>Sps<sub>1-3</sub> in the core to Alm<sub>56-58</sub>Prp<sub>22-25</sub>Grs<sub>18-19</sub>Sps<sub>0-1</sub> in the rim (Fig. 7b, Table 2). Scarce garnet fractures are filled by chlorite. Chlorite also replaces some of the garnet rims (Fig. 6c). Sodic amphibole classified as glaucophane is the main matrix phase (Fig. 6a-e). It shows compositional variations between core and rim (Fig. 7d, e, Table 2). The inner part has X<sub>Fe2+</sub> varying from 0.18 to 0.22, and then it increases in the rim to 0.30-0.35 (Fig. 7d, Table 2). Two types of white mica (phengite and paragonite) were identified in the blueschist. Phengite forms mainly elongate flakes in the matrix but also occurs as inclusions in the garnet porphyroblasts (Fig. 6c, d). It contains Si ranging from 3.34 to 3.45 apfu (Table 2). Paragonite ( $X_{Na} = 0.85 - 0.87$ ) forms polyphase inclusions in garnet (Fig. 6f, g). Epidote is present in the matrix (Fig. 6d) as well as inclusions



**Fig. 4** Photomicrographs and back-scattered electron (BSE) images for eclogite: **a** general overview: garnet porphyroblast surrounded by omphacitic matrix. Glaucophane, epidote, omphacite and rutile form inclusions in garnet. Chlorite fills the veins; **b**, **c** euhedral glaucophane grains within the matrix omphacite, note the chemical

zonation in BSE image; **d**, **e** oscillatory zoned omphacite grains; **f** phengite inclusion within garnet-I; **g** epidote inclusions in garnet-I; **h** prograde chlorite occurring in the core of garnet-I; **i** late carbonate veins cutting matrix omphacite

in garnet (Fig. 6f–h). It is characterized by  $X_{Fe3+}$  varying from 0.20 to 0.24 (Table 2). Chloritoid shows rather homogenous chemical composition with  $X_{Fe2+}$  around 0.35. Polyphase inclusions within garnet, sometimes with a tabular shape, consist of paragonite + kyanite and epidote + paragonite + chlorite. Such inclusions may be evidence for former presence of lawsonite (Fig. 4f–h; e.g. Tsujimori and Ernst 2014; Orozbaev et al. 2015). Carbonate forms scarce inclusions in the garnet or occurs in the matrix.



**Fig. 5** Chemistry of a major minerals in eclogite. **a** Mg, Fe, Ca and Mn compositional maps of garnet, warmer colors indicate higher concentration; **b** garnet major elements compositional profile along the transect marked on the Mg map; **c** Lu profile performed by LA-ICP-MS for the same garnet grain; **d** classification diagram of clino-

pyroxene following Morimoto et al. (1988) nomenclature; **e** plot of clinopyroxene compositions showing  $X_{Ae}$  and  $X_{Jd}$  end-members; **f** plot showing the range of sodic amphibole composition and compositional map of Fe for zoned glaucophane grain from the matrix

#### **Pressure-temperature estimates**

We used quartz in garnet (QuiG) Raman geobarometry coupled with Zr-in-rutile trace element thermobarometry to estimate the garnet growth conditions for the two studied HP lithologies. We also applied conventional thermobarometry methods using garnet–clinopyroxene–white mica to estimate eclogite crystallization conditions.

#### **Conventional geothermobarometry**

The coupled method of garnet-clinopyroxene  $Fe^{2+}$ -Mg exchange thermometer (Ravna 2000) and net-transfer

reaction 6diopside + 3muscovite = 3celadonite + 2grossular + pyrope geobarometer (Ravna and Terry 2004) have been applied to estimate the P–T conditions for eclogite sample KK14-55. The following activity models have been used: (1) the phengite activity model and solid solution of Holland and Powell (1998), (2) the clinopyroxene activity model of Holland (1990) and (3) the garnet activity model of Ganguly et al. (1996). The uncertainty of  $\pm$  60 °C is estimated for the calculated temperature and 0.3 GPa for pressure (Ravna and Terry 2004).

Three evolutionary stages were distinguished for eclogite based on the textures and mineral chemistry: M1, M2, and M3. Garnet-I core composition was coupled with phengite

Table 1 R	presentative anal	yses of ma	jor minera	uls for the e	sclogite										
Mineral	Grt <sup>M3</sup>	Grt <sup>M2</sup>	Grt	Grt <sup>MI</sup>	Grt	Grt	Ph <sup>MI</sup>	Ph <sup>M2</sup>	$Ph^{M3}$	Omph-I <sup>M1</sup>	Omph-I <sup>M2</sup>	Omph-II <sup>M3</sup>	Omph-II	Gln	Gln
Text.type	grt-II outerrim	grt-II	grt-I	grt-I	grt-∏	grt-II	incl grt-I	incl grt-Ⅱ	matrix	incl grt-I	incl grt-II	incl grt-II outerrim	matrix	core	rim
Analysis	15	23	67	89	103	115	8	7	3	16	61	20	22	2	5
$SiO_2$	37.91	38.22	37.96	37.94	38.80	38.37	52.03	52.28	49.53	55.96	57.06	56.76	56.03	58.65	56.57
$TiO_2$	0.30	0.30	0.08	0.07	0.10	0.10	0.20	0.13	0.09	0.08	0.08	0.05	0.17	0.01	0.19
$Al_2O_3$	20.63	20.89	21.31	21.27	21.74	22.03	25.87	24.79	26.12	9.05	11.52	11.22	9.76	10.71	8.00
$Cr_2O_3$	0.07	< 0.05	0.06	< 0.05	< 0.05	< 0.05	< 0.05	0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05
FeO	29.21	30.24	32.83	32.53	28.61	25.97	3.62	3.83	4.12	8.03	6.00	8.00	6.58	90.6	16.82
MnO	0.24	0.04	0.98	0.22	0.32	0.72	< 0.04	< 0.04	< 0.04	0.07	0.04	< 0.04	0.09	0.07	0.05
MgO	4.14	3.28	4.60	4.33	4.25	5.87	3.73	3.69	2.79	7.14	7.05	6.70	7.85	12.05	7.88
CaO	6.90	7.03	1.69	3.71	6.91	6.00	0.02	0.01	0.03	11.29	10.63	10.56	11.79	0.79	0.89
$Na_2O$	0.10	0.05	0.16	0.08	0.07	0.04	0.32	0.18	0.32	7.52	8.11	8.26	7.29	7.41	7.06
$K_2O$	< 0.04	< 0.04	< 0.04	< 0.04	< 0.04	< 0.04	10.34	10.49	10.22	< 0.04	< 0.04	< 0.04	< 0.04	< 0.04	< 0.04
BaO	na	na	na	na	na	na	0.40	0.28	0.35	< 0.04	< 0.04	< 0.04	< 0.04	< 0.04	< 0.04
Total	99.50	100.04	99.67	100.18	100.81	99.11	96.53	95.72	93.61	99.13	100.49	101.57	99.59	98.77	97.47
Si	3.010	3.025	3.019	3.005	3.019	3.002	3.453	3.472	3.405	2.021	2.013	1.991	2.006	7.902	7.992
Ï	0.018	0.018	0.005	0.004	0.006	0.006	0.010	0.007	0.005	0.002	0.002	0.001	0.005	0.001	0.020
AI	1.931	1.949	1.998	1.986	1.994	2.032	2.023	1.941	2.116	0.385	0.479	0.464	0.412	1.700	1.332
Cr	0.004	0.000	0.004	0.001	0.000	0.000	0.000	0.002	0.000	0.000	0.000	0.000	0.001	0.000	0.000
$\mathrm{Fe}^{3+}$	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.095	0.045	0.113	0.073	0.205	0.438
$\mathrm{Fe}^{2+}$	1.940	2.002	2.183	2.155	1.862	1.699	0.201	0.213	0.237	0.147	0.132	0.121	0.124	0.815	1.550
Mn	0.016	0.002	0.066	0.015	0.021	0.048	0.000	0.000	0.002	0.002	0.001	0.000	0.003	0.008	0.000
Mg	0.490	0.387	0.545	0.511	0.493	0.684	0.369	0.365	0.286	0.384	0.371	0.350	0.419	2.420	1.660
Ca	0.587	0.596	0.144	0.315	0.576	0.503	0.001	0.000	0.002	0.437	0.402	0.397	0.452	0.114	0.134
Na	0.015	0.007	0.024	0.012	0.011	0.007	0.041	0.023	0.043	0.526	0.555	0.562	0.506	1.936	1.934
K	0.000	0.000	0.001	0.000	0.000	0.000	0.875	0.889	0.896	0.000	0.000	0.000	0.000	0.003	0.003
Ba	na	na	na	na	na	na	0.011	0.007	0.010	0.000	0.000	0.000	0.000	0.000	0.000
Total	8.011	7.986	7.987	8.003	7.983	7.980	6.984	6.919	7.002	4.000	4.000	4.000	4.000	15.104	15.063
$\mathrm{X}_{\mathrm{Alm}}$	0.640	0.670	0.743	0.719	0.631	0.579			ЪЦ	38	51	45	40		
$\mathbf{X}_{\mathbf{Sps}}$	0.005	0.001	0.022	0.005	0.007	0.016			$\mathbf{X}_{\mathrm{Mg}}$	0.723	0.738	0.743	0.771		
$\mathbf{X}_{\mathrm{Prp}}$	0.161	0.129	0.185	0.171	0.167	0.233									
$\mathbf{X}_{\mathrm{Grs}}$	0.194	0.200	0.049	0.105	0.195	0.171									
$\mathbf{X}_{\mathrm{Fe}}$	0.798	0.838	0.800	0.808	0.791	0.713								0.252	0.483
$X_{Alm} = Fe^2$ % = (Na -	$^{+/(Fe^{2+} + Mg + C_{\delta})}$ Fe <sup>3+</sup> - (2 × Ti))	t+Mn); X, × 100%	sps = Mn/(	Fe <sup>2+</sup> + Mg	+Ca+Mı	$\eta$ ; $X_{Prp} = l$	Mg/(Fe <sup>2+</sup> +]	Mg+Ca+M	n); X <sub>Grs</sub> =	:Ca/(Fe <sup>2+</sup> + N	lg+Ca+Mn)	$X_{Fe} = Fe^{2+}/(Fe^{2+} + M)$	g); $X_{Mg} = N$	1g/(Fe <sup>2+</sup> +	Mg); Jd
Structural t	formulae calculate	ad on the h	asis of: 12	2 ox veens 1	or parnet.	23 oxvøei	is and Fe <sup>3+</sup> ;	as average af	ter Hollan	d and Blundv	(1994) for N	a-amphibole. 11 ox vge	ans for phene	oite and 6	OXVPENS
for clinopy	roxene (Morimote	0 1988)			2			-0					I was not		

inclusions from the garnet-I core and core of matrix omphacite for stage M1, Garnet mantle along with phengite from garnet mantle and the mantle part of matrix omphacite were used for estimating the M2 conditions, and garnet-II rim together with matrix phengite and the rim of omphacite from the matrix were combined to estimate M3 conditions (Table 1, Fig. 8a). The calculations yielded P–T conditions of  $1.6 \pm 0.3$  GPa at  $460 \pm 60$  °C for M1. The conventional geothermobarometry for M2 and M3 stages gave  $2.3 \pm 0.3$ GPa at  $507 \pm 60$  °C and  $2.1 \pm 0.3$  GPa at  $553 \pm 60$  °C, respectively (Fig. 8a).

#### Quartz-in-garnet (QuiG) barometry

Garnets from the eclogite and blueschist samples do not contain abundant quartz inclusions. In the eclogite sample, both garnet-I and garnet-II crystals rarely contain quartz inclusions, and they occur randomly distributed throughout garnet host crystals. Quartz inclusions in garnet crystals from the blueschist sample typically occurred in the cores of crystals. The inclusions we measured are approximately spherical, range from 3 to 30  $\mu$ m in diameter and are fully encapsulated by garnet hosts.

Eighteen quartz inclusions in garnet were measured in eclogite sample KK14-55, eleven of them are in garnet-I and seven are in garnet-II. The strained quartz inclusions have Raman peaks upshifted by  $\Delta 128$ ,  $\Delta 206$ , and  $\Delta 464 \text{ cm}^{-1}$  of  $4.70 \pm 0.12$ ,  $17.09 \pm 0.65$  and  $6.42 \pm 0.24 \text{ cm}^{-1}$  for garnet-I and  $4.82 \pm 0.59$ ,  $18.87 \pm 1.47$  and  $6.86 \pm 0.71 \text{ cm}^{-1}$  for garnet-II, which give average  $P_{incl}$  of  $0.76 \pm 0.03$  GPa (Fig. 8a, Kośmińska et al. 2023) for garnet-I and  $P_{incl}$  of  $0.86 \pm 0.05$  GPa for garnet-II (Fig. 8a, Kośmińska et al. 2023). The average isomekes plot at lower-pressure conditions than the results of conventional geothermobarometry and below the eclogite facies conditions (Fig. 8a). The maximum and minimum isomekes for garnet-I and garnet-II in Fig. 8a, b show the range of potential entrapment pressures (Fig. 8a).

Because inclusions were rare, we measured only 10 inclusions in garnets from the blueschist sample. The strained quartz inclusions have Raman peaks upshifted by  $\Delta 128$ ,  $\Delta 206$ , and  $\Delta 464 \text{ cm}^{-1}$  of  $5.07 \pm 0.21$ ,  $18.28 \pm 0.72$  and  $6.73 \pm 0.33 \text{ cm}^{-1}$  (Kośmińska et al. 2023). The results define one consistent quartz population which gives the average  $P_{incl}$  of  $0.82 \pm 0.03$  GPa (Fig. 8b). The minimum and maximum isomekes are marked by dotted lines, which indicate the potential entrapment pressures (Fig. 8b).

#### Zirconium-in-rutile thermometry

The eclogite sample KK14-55 contains rutile forming stubby, anhedral crystals up to  $500 \,\mu\text{m}$  in diameter. It forms inclusions within garnet-I and garnet-II and occurs in the omphacite-dominated matrix. The Zr concentrations range

from 36 to 60 ppm for rutile inclusions in the garnet-I (average 45 ppm, Kośmińska et al. 2023), while Zr in rutile inclusions from garnet-II varies from 48 to 74 ppm (average 62 ppm, Kośmińska et al. 2023). Grains in the matrix contain between 40 and 81 ppm of Zr (average 64 ppm). Zirconium-in-rutile temperatures have been estimated using pressure sensitive calibration of Tomkins et al. (2007). The equation for the  $\alpha$ -quartz stability field has been used. The Zr-in-rutile thermometry was applied to rutile inclusions in the garnet and rutile from the matrix. The uncertainty of  $\pm$  30 °C is estimated combining the analytical error and the calibration. The pressure obtained by garnet-phengite-clinopyroxene thermobarometry was used for the P-T calculations of the growth of garnet-I (M1) and garnet-II (M2). The measured Zr concentrations correspond to the temperature of  $500 \pm 30$  °C (at fixed pressure of 1.8 GPa) for the inclusions from garnet-I and  $550 \pm 30$  °C (at fixed pressure of 2.3 GPa) for inclusions from garnet-II and matrix rutile (Fig. 8a). A similar approach was applied to the blueschist sample KK12-12E. Rutile forms stubby, anhedral crystals that are considerably smaller than in the eclogite (up to 100 µm in diameter). It occurs as inclusions in garnet and within the matrix. Zr content ranges from 24 to 78 ppm for the rutile inclusions in garnet (average 35 ppm), and 50–94 ppm for the matrix rutile (average 73 ppm). The calculations yield temperatures of  $500 \pm 30$  °C for the inclusions and  $540 \pm 30$  °C (at fixed pressure of 2.0 GPa) for rutile included in the glaucophanitic matrix (Fig. 8b). The combined QuiG barometry and Zr-in-rutile thermometry indicate that the conditions of the garnet growth reached up to  $2.1 \pm 0.03$  GPa at  $500 \pm 30$  °C (garnet-I) and  $2.25 \pm 0.05$ GPa at  $550 \pm 30$  °C (garnet-II) for the eclogite, and peakpressure conditions at  $2.0 \pm 0.03$  GPa at  $500 \pm 30$  °C for the blueschist (Fig. 8).

# Zircon U–Pb geochronology and trace element analyses

Zircon in eclogite sample KK14-55 occurs in the omphacitic matrix and as inclusions in garnet. It is not very abundant and forms mostly tiny ( $<70 \ \mu m$  in the longest dimension), subhedral, prismatic grains. Bigger grains ( $>100 \ \mu m$ ) are rare (Fig. 9). Zircon contains inclusions of quartz, rutile and white mica and is complex, typically showing distinctive core and rim domains (Fig. 9). CL images reveal oscillatory zoning within some grains (Fig. 9f, Supplementary materials S2, e.g., zircon 31, 33, 39) but no zoning in the remaining grains. Zircon rims are observed on grains within garnet (e.g., zircon 29, Fig. 9e) as well as in the matrix.

The analyses define a discordant array with <sup>206</sup>Pb/<sup>238</sup>U dates ranging from 1655 to 363 Ma that reflect the combined effects of xenocrystic cores of variable age, Pb-loss and common Pb (Fig. 9a, Kośmińska et al. 2023). The



◄Fig. 6 Photomicrographs and back-scattered electron (BSE) images for blueschist: a general overview: garnet porphyroblast surrounded by glaucophanitic matrix; b chloritoid together with epidote occurring between glaucophane and garnet; c garnet partially replaced by chlorite; d epidote and phengite forming aggregates within glaucophane; e BSE image showing zoned glaucophane in the matrix; f-h polymineral inclusions from the garnet core composed of: epidote + paragonite + quartz + ilmenite ± kyanite

cores show steep HREE, Th/U = 0.02-0.80 and a negative Eu anomaly that are characteristic of igneous zircon (Fig. 9c). Regression of 12 younger rim analyses uncorrected for common Pb defines a lower intercept of  $480 \pm 12$  Ma (MSWD = 1.0, Fig. 9a). This estimate is consistent with a weighted mean  $^{206}$ Pb/ $^{238}$ U age of  $482 \pm 10$  Ma (MSWD = 1.1, Fig. 9b, e) provided by 11 of the rim analyses with lower common Pb contents. The two youngest analyses (spots 8.1 and 9.1, Fig. 9, Kośmińska et al. 2023) are interpreted to record Pb loss. The rim analyses are characterized by low Th/U ratio (0.01–0.02) and a small Eu anomaly (Eu/Eu\* > 40, Fig. 9c, d, Kośmińska et al. 2023). The rim analyses have similar steep HREE patterns but slightly lower abundances when compared with zircon core results.

# Monazite U–Pb geochronology and trace element analyses

Monazite in the eclogite forms rare, tiny inclusions  $(<30 \ \mu\text{m})$  in garnet-II or occurs in the matrix included in the allanite-epidote aggregates (Fig. 10). It is commonly surrounded by a corona composed of apatite and allanite (Fig. 10c–f). Monazite is very weakly zoned (Fig. 10c–f). Y for example varies from 13,457 to 23,037 ppm. Monazite inclusions in garnet are characterized by lower Th (9005–9533 ppm) and U (435–621 ppm) than matrix monazite characterized by Th between 12,971 and 20,555 ppm and U ranging from 1727 to 2757 ppm. The REE patterns show enrichment in LREE, slight depletion in HREE and a moderate negative Eu anomaly (Fig. 12g, Kośmińska et al. 2023). Seven analyses define a <sup>206</sup>Pb/<sup>238</sup>U weighted mean age of 471±6 Ma (MSWD=1.4, Fig. 12h, Kośmińska et al. 2023).

Monazite in the blueschist forms very few, small (< 30  $\mu$ m) grains in the glaucophane-rich matrix. One statistically younger analysis was not included (analysis 3.1, Kośmińska et al. 2023). Monazite is Th rich and shows very weak Y zonation (Fig. 11a–c, Kośmińska et al. 2023). The trace element signatures display enrichment in LREE, depletion in HREE and a flat Eu anomaly (Fig. 11d, Kośmińska et al. 2023). Three analyses define a weighted mean <sup>206</sup>Pb/<sup>238</sup>U age of 486±6 Ma (MSWD=0.3, Fig. 11e, Kośmińska et al. 2023).

# Allanite U–Pb geochronology and trace element analyses

Only allanite surrounding monazite in the matrix was dated (Fig. 10c, d), since allanite inclusions in garnet were smaller than the analytical spot. Thirteen analyses give a data-defined 3-D total Pb/U isochron with a lower intercept of  $392 \pm 50$  Ma (MSWD = 1.6, Kośmińska et al. 2023, supplementary materials S3). The trace element pattern shows enrichments in LREE relative to HREE and a negative Eu anomaly (Fig. 10g, Kośmińska et al. 2023).

# LA-ICPMS analyses of garnet

The Lu profiles of garnet porphyroblasts obtained by LA-ICPMS (Fig. 5c) run through one of the garnet grains of sample KK14-55 measured with electron microprobe (Fig. 5a, b). As already observed in the major element distribution, the garnet can be divided into garnet-I and garnet-II according to the Lu concentration. Garnet-I shows a bell-shaped distribution, typical for prograde growth zoning, and garnet-II displays either a generally continuous increase towards the outer rim or oscillatory zoning.

A Lu abundance profile of sample KK12-12E through one garnet grain is presented in Fig. 7c. It clearly shows a bell-shaped distribution similar to the major element pattern.

#### Garnet Lu–Hf geochronology

A total of five garnet separates, one clinopyroxene separate and two whole rock splits were analyzed from sample KK14-55 (Table 3). The bomb-digested whole rock split does not fit on an isochron with the garnet separates and the tabletop-digested whole rock split. This is interpreted to reflect the presence of Hf-rich phases inherited from older geological processes unrelated to high-pressure metamorphism recorded by the garnets. Therefore, a clinopyroxene separate was analyzed, as clinopyroxene is part of the high-pressure paragenesis. This clinopyroxene, the tabletop-digested whole rock split, and four of the garnet separates fall on an isochron giving an age of  $467 \pm 3$  Ma (MSWD = 3.3, n = 6; Fig. 12a). The fifth garnet separate falls slightly below this isochron and together with the clinopyroxene separate and the whole rock gives a younger age of  $458 \pm 3$  Ma (MSWD = 0.6, n = 3; Fig. 12b). Since all garnet separates have roughly the same Lu content (10–12 ppm), they are all inferred to represent equal mixtures of garnet-I and garnet-II.

Lu–Hf isotopic analyses are shown in Table 3. Four garnet separates, three table-top digested whole rock splits, one bomb-digested whole rock split and two glaucophane separates were analyzed for sample KK12-12E. All analyzed separates and whole rock splits fit define an isochron with an age of  $471 \pm 4$  Ma (MSWD=2.8, n=10, Fig. 12c).



**Fig. 7** Chemistry of garnet and sodic amphibole in blueschist. **a** Mg, Fe, Ca and Mn compositional maps of garnet, warmer colors indicate higher concentration; **b** garnet major elements compositional profile along the transect marked on the Mg map; **c** Lu profile performed by

# Discussion

# P–T evolution of the Upper unit, Vestgötabreen complex

HP-LT rocks from the Upper unit of the Vestgötabreen Complex show a clockwise P–T path reaching peak-pressure conditions up to  $2.3 \pm 0.3$  GPa at  $507 \pm 60$  °C (eclogite) and  $2.0 \pm 0.03$  GPa at  $500 \pm 30$  °C (blueschist) that are typical for HP-LT subduction regimes (Evans 1990; Agard et al. 2009; Tsujimori and Ernst 2014). The peak metamorphic estimates for eclogite give similar pressure, but slightly lower temperature than those conditions estimated by Hirajima et al. (1988), whereas the peak conditions for blueschists agree with the Nordenskiöld Land blueschists that are also interpreted to be a part of the Vestgötabreen Complex (Kośmińska et al. 2014).

The zonation of major minerals combined with available thermobarometers and abundant accessory minerals allows

LA-ICP-MS for the same garnet grain;  $\mathbf{d}$  plot showing the composition of sodic amphibole;  $\mathbf{e}$  compositional map of Fe showing zoned glaucophane in the matrix

for detailed P-T path reconstructions that include three evolutionary stages: M1, M2 and M3. Stage M1 in eclogite is interpreted as prograde metamorphism at  $1.6 \pm 0.3$  GPa and  $460 \pm 60$  °C (Fig. 13) represented by the mineral assemblage of garnet-I, omphacite-I, and epidote, glaucophane, lawsonite and phengite inclusions in the garnet that have grown under blueschist facies conditions (Figs. 4a, 5). P-T conditions for M1 were estimated based on conventional geothermobarometry combined with QuiG barometry for quartz inclusions within garnet-I and Zr-in-rutile in garnet-I. Stage M2 represents the peak pressure conditions and incipient decompression under lawsonite-eclogite facies conditions at  $2.3 \pm 0.3$  GPa and  $507 \pm 60$  °C (Fig. 13), based on conventional geothermobarometry coupled with QuiG barometry for inclusions in garnet-II and Zr-in-rutile thermometry for the rutile inclusions in garnet-II and matrix. The peak temperature condition (M3) is  $553 \pm 60$  °C at  $2.1 \pm 0.3$  GPa based on conventional geothermobarometry combined with Zr-in-rutile in the matrix-located rutile. The last stage of the

Table 2 Re <sub>l</sub>	presentative	analyses o	of major m	inerals for t	he bluesch	ist											
Mineral	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Ph	Ph	Ph	Ph	Pg	Gln	Gln	Gln	Gln	Gln
Text.type	rim	rim	core	core	core	rim	rim									rim	rim
Analysis	1	16	26	31	38	06	98	1	2	5	14	9	1	9	12	7	8
SiO <sub>2</sub>	38.83	38.01	37.46	37.67	36.92	38.13	38.40	50.92	50.86	51.60	52.23	46.13	56.58	56.51	56.64	58.41	58.39
$TiO_2$	0.09	0.10	0.09	0.12	1.71	0.07	0.03	0.26	0.34	0.18	0.13	0.04	0.07	0.00	0.06	0.03	0.02
$Al_2O_3$	21.44	21.07	20.75	20.90	20.45	21.19	21.62	27.80	27.40	26.88	25.62	38.49	9.97	9.71	9.88	9.43	8.47
$Cr_2O_3$	< 0.05	< 0.05	0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05
FeO	26.26	29.27	29.57	28.94	29.45	29.73	29.58	3.71	3.41	3.90	4.05	0.95	8.48	9.64	8.27	10.09	13.29
MnO	0.44	0.73	1.45	1.54	1.35	0.21	0.21	< 0.04	< 0.04	< 0.04	< 0.04	< 0.04	12.93	12.02	12.30	11.99	9.87
MgO	6.58	3.36	2.51	2.53	2.51	6.63	8.19	2.99	3.14	3.05	3.27	0.10	0.00	0.01	0.03	0.03	0.03
CaO	6.49	7.50	7.89	8.02	7.26	3.71	1.85	0.00	0.02	0.01	0.00	0.23	0.71	0.74	0.71	1.02	0.43
$Na_2O$	< 0.04	< 0.04	< 0.04	< 0.04	0.07	< 0.04	< 0.04	0.82	0.89	0.43	0.35	6.81	7.43	7.43	7.23	5.93	6.28
$K_2O$	< 0.04	< 0.04	< 0.04	< 0.04	< 0.04	< 0.04	< 0.04	9.51	9.60	10.02	10.35	1.40	0.05	< 0.04	< 0.04	< 0.04	< 0.04
Total	100.14	100.09	77.66	99.78	99.74	99.68	06.66	96.02	95.66	96.14	96.03	94.24	96.22	96.08	95.16	96.92	96.78
Si	3.010	3.008	2.997	3.005	2.954	2.996	2.988	3.341	3.356	3.390	3.451	2.991	7.806	7.841	7.893	8.039	8.098
Ti	0.005	0.006	0.005	0.007	0.103	0.004	0.002	0.013	0.017	0.014	0.007	0.003	0.007	0.000	0.006	0.003	0.002
AI	1.959	1.966	1.958	1.966	1.930	1.962	1.984	2.150	2.131	2.083	1.995	2.941	1.621	1.589	1.623	1.531	1.384
Cr	0.000	0.002	0.003	0.002	0.001	0.001	0.002	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000
$\mathrm{Fe}^{3+}$								0.000	0.000	0.000	0.000	0.000	0.400	0.435	0.318	0.111	0.463
$\mathrm{Fe}^{2+}$	1.702	1.937	1.979	1.931	1.971	1.953	1.925	0.203	0.188	0.188	0.224	0.112	0.578	0.685	0.646	1.050	1.079
Mn	0.029	0.049	0.098	0.104	0.092	0.014	0.014	0.001	0.000	0.002	0.001	0.000	0.000	0.001	0.003	0.003	0.004
Mg	0.759	0.396	0.299	0.301	0.299	0.775	0.949	0.293	0.308	0.323	0.322	0.161	2.659	2.485	2.555	2.459	2.039
Ca	0.539	0.636	0.677	0.686	0.623	0.312	0.154	0.000	0.002	0.000	0.000	0.000	0.105	0.109	0.107	0.151	0.064
Na	0.003	0.002	0.001	0.004	0.011	0.003	0.001	0.104	0.114	0.099	0.044	0.022	1.989	1.999	1.954	1.581	1.689
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.796	0.808	0.811	0.872	0.436	0.009	0.004	0.005	0.000	0.000
Total	8.007	8.003	8.018	8.006	7.983	8.020	8.018	6.900	6.923	6.911	6.916	6.668	15.175	15.149	15.110	14.928	14.821
$\mathbf{X}_{\mathrm{Alm}}$	0.562	0.642	0.648	0.639	0.660	0.639	0.633										
$\mathbf{X}_{\mathrm{Sps}}$	0.00	0.016	0.032	0.034	0.031	0.005	0.005										
$\mathbf{X}_{\mathrm{Prp}}$	0.251	0.131	0.098	0.100	0.100	0.254	0.312										
$\mathbf{X}_{\mathrm{Grs}}$	0.178	0.211	0.222	0.227	0.209	0.102	0.051										
$\mathbf{X}_{\mathrm{Fe}}$	0.692	0.830	0.869	0.865	0.868	0.716	0.670						0.179	0.216	0.202	0.299	0.346
X <sub>Alm</sub> = Fe/(I Structural fc	Te + Mg + C	a + Mn); X sulated on t	(Sps=Mn/( the basis o	Fe + Mg + ( f: 12 oxyge	Ca+Mn); 7 ns for garn	X <sub>Prp</sub> =Mg/( let, 23 oxy;	(Fe + Mg + zens and Fe	Ca + Mn); ] <sup>3+</sup> as avera	X <sub>Grs</sub> =Ca/( ge after Ho	Fe+Mg+ olland and	Ca+Mn); Blundy (19	X <sub>Fe</sub> = Fe/(F 94) for Na-	e + Mg) amphibol	e, and 11 o	tygens for 1	bhengite	



**Fig.8** Summary diagram showing pressure-temperature (P–T) estimates for eclogite (**a**) and blueschist (**b**). **a** Summary P–T diagram for eclogite. Diagram showing QuiG results for garnet-I and garnet-II. Solid lines indicate the average isomekes for QuiG results, whereas dotted lines are minimum and maximum isomekes. Lighter green color refers to the results for the garnet-I, whereas darker marks result for garnet-II. Crosses are indicating the results of Grt–Ph–Cpx geothermobarometry for different textural positions. Zr-in-rutile results

are shown for inclusions in garnet-I (lighter color), garnet-II (darker green color) and measurements for the matrix rutile (transparent with pattern). QuiG results are marked as shades along the P–T path for simplicity. **b** P–T results for blueschist. Average (solid line) and minimum/maximum (dotted lines) isomekes are shown for QuiG results. QuiG is coupled with Zr-in-rutile thermometry for inclusions (lighter blue color) and matrix rutile grains (transparent with pattern). Metamorphic facies and their abbreviations are from Liou et al. (2004)

P–T path is recorded under lower-T conditions, most probably in greenschist facies, as demonstrated by chloritization of garnet.

Garnet-II is Ca-enriched and the chemical difference within a single porphyroblast is very sharp (Fig. 5a, b). The distinctive zonation is suggesting Ca influx, which contributed to the growth of garnet-II. We suggest that the garnet growth was prolonged, and the change of chemistry is caused by the break-down of another Ca-rich phase. The most likely source of Ca may be lawsonite or epidote (e.g. Fassmer et al. 2016). Lawsonite inclusions were not found in the studied sample, however Hirajima et al. (1988) reported very scarce, minute lawsonite inclusions from the cores of the garnet in eclogites. The potential break-down of lawsonite may have caused the formation of garnet-II and omphacite-II, following the reaction proposed by Groppo and Castelli (2010): Lws + Qz + Chl + Omp = Grt + Omp + H<sub>2</sub>O. Moreover, garnet-II is enriched in Lu (Fig. 5c), which can also reflect lawsonite decomposition since it is a sink of REE (e.g. Tribuzio et al. 1996; Spandler et al. 2003; Mulcahy et al. 2009).

Garnet shows oscillatory zonation in the outermost part of garnet-II, which is clearly visible on the Mn compositional map (Fig. 5a). Such a phenomenon can be interpreted to be the result of fluctuations in P-T conditions (e.g. García-Casco et al. 2002; Kohn 2004; Li et al. 2016; Viete et al. 2018) or influx of external fluids (e.g. Yardley et al. 1991; Angiboust et al. 2011; Martin et al. 2011). It is apparent that the Vestgötabreen rocks experienced massive dehydration of phases that are typical for LT blueschist facies. This process must have caused an increase of fluid pressure and could have contributed to local P fluctuation, which in turn could have triggered the dissolution-reprecipitation of the outermost garnet-II now expressed as an oscillatory Mn pattern. Garnet-I, omphacite and epidote also exhibit oscillatory zonation (Figs. 4e, g, 5a). All of these observations suggest that studied eclogite experienced several fluid infiltration events. However, it is hard to assess whether the fluid was internal or with an addition of external influx based on the data presented in this study. Nonetheless, it is apparent that several fluid infiltration events could have affected the studied unit and caused the aforementioned oscillatory zoning



**Fig.9 a** Tera-Wasserburg plot of sensitive high-resolution ion microprobe reverse-geometry (SHRIMP-RG) U–Pb analyses for zircons from eclogite, dark blue analyses were used for constraining the age (see text for more details), U–Pb data are plotted as  $2\sigma$  error ellipses uncorrected for common Pb, MSWD—mean square of the weighted deviates; **b** weighted average age calculated for the younger zircon

analyses; **c** chondrite-normalized REE plot; **d** Th/U vs. Eu\* plot; **e** back-scattered electron (BSE) images with analytical spots marked for the younger analyses which were used for age constrains; **f** cathodoluminescence (CL) images of representative zircon grains analyzed in the eclogite

features associated with fluid-aided potential dissolution and reprecipitation of several rock forming minerals including garnet via general scheme proposed by Putnis (2002).

The combination of QuiG barometry and Zr-in-rutile thermobarometry for the blueschist sample, KK12-12E, gives peak pressure conditions of ca.  $2.0 \pm 0.03$  GPa at  $500 \pm 30$  °C. Garnet in the blueschist contains polymineral inclusions built of paragonite + kyanite + epidote (Fig. 6f, h), and epidote + paragonite (Fig. 6g), occasionally resembling a tabular shape. The aggregates of epidote + paragonite occur also in the matrix (Fig. 6d). These textures suggest the former presence of lawsonite (e.g. Poli and Schmidt 2004; Tsujimori and Ernst 2014; Orozbaev et al. 2015). The lawsonite break down reaction could have happened along the retrograde path according to the reaction(s): Grt + Gln + Lws = Ep + Pg + H<sub>2</sub>O; Lws + Ab = Pg + Ep + Qz + H<sub>2</sub>O; Lws = Ep + Ky + Qz + H<sub>2</sub>O. Minor retrogression features, such as chlorite replacing garnet, suggest a late overprint at greenschist facies conditions.

# **QuiG barometry**

Recent experiments show that changing pressure can partially reset quartz inclusions in garnet (Pummell and Thomas 2022). Experimental results at 700 and 800 °C show that quartz inclusions in garnet reset to pressures between the entrapment pressure and the final pressure. Natural samples also show evidence of partial resetting



Fig. 10 a, b Back-scattered electron (BSE) images of allanite aggregates occurring in the omphacitic matrix of eclogite; c-f BSE images and compositional maps for dated monazite in the eclogite, black ellipses indicate the analytical spot for monazite and grey ellipses for

allanite; **g** chondrite-normalized REE plot for monazite and allanite of sensitive high-resolution ion microprobe reverse-geometry (SHRIMP-RG) trace elements analyses; **h** weighted average age calculated for the analyzed monazite grains

of quartz inclusions in garnet. Quartz inclusions in garnet from natural rocks give isomekes that span a large range of pressures, and some of the isomekes are lower than interpreted peak pressure conditions (Kośmińska et al. 2020; Wolfe and Spear 2020; Simon et al. 2022). At any given crystallization temperature, QuiG gives a distribution of entrapment pressures. Quartz inclusions in garnet that have the highest  $P_{incl}$  give isomekes that are commonly similar to the peak P conditions estimated by other methods. Numerous studies interpret the highest-P inclusions to give isomekes most representative of the peak metamorphic conditions (Spear et al. 2014; Castro and Spear 2017; Wolfe and Spear 2018, 2020, 2022; Thomas and Spear 2018; Kośmińska et al. 2020; Spear and Wolfe 2020). Here, the temperature obtained from the combination of Zr-in-rutile and garnet-phengite-clinopyroxene thermobarometry for the eclogite and Zr-in-rutile thermometry for the blueschist were used to calculate entrapment pressures. The eclogite sample shows a spread of pressures between  $1.3 \pm 0.03$  and  $2.3 \pm 0.05$  GPa (Fig. 8a), Fig. 11 a-c Back-scattered electron (BSE) images and Th, Y compositional maps for dated monazite in the blueschist; d chondrite-normalized REE plot of sensitive high-resolution ion microprobe reverse-geometry (SHRIMP-RG) trace elements analyses; e weighted average age calculated for the analyzed monazite grains



Table 3Lu and Hf isotopiccompositions of whole rocks(wr), garnet separates (Grt),clinopyroxene separates (Cpx)and glaucophane separates(Gln)

Mineral	Lu (ppm)	Hf (ppm)	<sup>176</sup> Lu/ <sup>177</sup> Hf	Error	<sup>176</sup> Hf/ <sup>177</sup> Hf	Error
Eclogite sa	mple KK14-55					
wr tt	2.29	1.85	0.1757	0.0004	0.284278	1.2E-05
wr b	2.42	9.99	0.03441	0.00007	0.282992	1.4E-05
Срх	0.679	1.06	0.09112	0.00018	0.283528	6.2E-05
Grt 1	10.1	1.37	1.050	0.002	0.291912	1.3E-05
Grt 2	11.7	1.10	1.509	0.003	0.295714	7.6E-05
Grt 3	11.1	1.43	1.103	0.002	0.292424	2.3E-05
Grt 4	11.4	1.27	1.276	0.003	0.293992	5.5E-05
Grt 5	10.7	1.04	1.463	0.003	0.295542	7.8E-05
Blueschist	sample KK12-12	2E				
wr tt 1	0.477	0.164	0.4122	0.0008	0.286485	4.9E-05
wr tt 2	0.459	0.110	0.5928	0.0012	0.288066	4.7E-05
wr tt 3	0.845	0.211	0.5703	0.0011	0.287911	4.9E-05
wr b	0.534	3.47	0.02181	0.00004	0.282963	1.3E-05
Gln 1	0.0501	0.0565	0.1259	0.0003	0.283936	8.1E-05
Gln 2	0.0515	0.0870	0.08401	0.00017	0.283525	0.000080
Grt 1	0.632	0.106	0.8498	0.0017	0.290246	3.8E-05
Grt 2	0.707	0.102	0.9823	0.002	0.291463	0.00015
Grt 3	0.744	0.108	0.9795	0.002	0.291427	5.4E-05
Grt 4	0.752	0.103	1.034	0.002	0.291794	0.00013

whereas blueschist shows a smaller spread of  $1.7 \pm 0.03$  to  $2.0 \pm 0.03$  GPa (Fig. 8b), thus suggesting a different degree

of resetting. Based on the observed mineral assemblages and other equilibrium-based techniques, we decided to use



Fig. 12 Lu–Hf isochrones for the eclogite (a, b) and blueschist (c) samples.  $2\sigma$  uncertainties are used. Calculated initial values and ages are based on  $\lambda^{176}$ Lu = 1.865 × 10<sup>-11</sup> year<sup>-1</sup> (Scherer et al. 2001)



**Fig. 13** Summary diagram showing the P–T–t evolution of the Vestgötabreen Complex. Blue dotted line represents P–T path for the blueschist, whereas green dotted line for the eclogite. Black line shows P–T path obtained by Hirajima et al. (1988). Stages M1 (prograde metamorphism), M2 (peak-P conditions) and M3 (peak-T conditions were distinguished for the eclogite sample, whereas only peak-P conditions were deciphered for blueschist. The ages refer to

the maximum isomeke as the best estimate of the peak metamorphic conditions.

# Interpretation of zircon ages

The small size of zircon grains as well as very complex CL zoning patterns make interpretation of the results difficult.

the results of geochronology presented in this paper (U–Pb zircon, U–Pb monazite, Lu–Hf garnet). P–T estimates: [1] Nordenskiöld Land blueschist (Kośmińska et al. 2014), [2] the Vestgötabreen Complex Upper unit (Hirajima et al. 1988), [3] the Vestgötabreen Complex Lower unit (Agard et al. 2005). Metamorphic facies and their abbreviations are from Liou et al. (2004). Florencite-out reaction (marked as mnz-in) is after Barthélemy and Chopin (2006)

The zircon cores give U–Pb dates ranging from the Paleoproterozoic to the Cambrian. The spread in dates combined with oscillatory zoning (Fig. 9) and trace element patterns (Fig. 9c, d) typical of igneous zircon (i.e. Corfu et al. 2003; Hoskin and Schaltegger 2003) suggest the cores are xenocrysts in the mafic protolith. The xenocryst ages are compatible with detrital components observed in sedimentary units of the Southwestern province (e.g., Ziemniak et al. 2019). The early Ordovician <sup>206</sup>Pb/<sup>238</sup>U weighted mean age of  $482 \pm 10$  Ma obtained for zircon rims is interpreted to record metamorphic growth (Fig. 9b). The presence of rims on zircon included within garnet suggests that at least some metamorphic overgrowth occurred before final garnet formation. The trace element patterns (Fig. 9c, d) of zircon rims display steep HREE patterns that are similar to the cores, but have a less pronounced negative Eu anomaly. Thus, we infer that zircon growth occurred along the prograde path accompanied by no or limited garnet growth (Rubatto 2002). The Early Ordovician age for zircon rims is in agreement with U–Pb zircon results of Bernard-Griffiths et al. (1993).

#### Monazite precursor and growth conditions

Monazite is common in eclogite-facies rocks, but very scarce in metamorphosed mafic and ultramafic rocks (e.g. Schulz 2021). One of the very rare places where monazite has been described and dated in eclogite is the Colorado Plateau (Schulze et al. 2015). The growth of monazite in the eclogite xenoliths was suggested to be metasomatic, caused by hydrothermal fluids linked to the Great Hydration event. The metasomatism driven growth of monazite was also suggested for mafic and ultramafic bodies from the Dabie Mountains, China (Liou and Zhang 1998; Ayers et al. 2002; Jahn et al. 2003). Philippot and Selverstone (1991) described monazite as a daughter mineral in fluid inclusions in the eclogitic veins from the Monviso ophiolite in the Alps and interpreted its origin as precipitation from a hydrous fluid.

Lawsonite is a potential source of LREE elements in HP-LT rocks (e.g. Tribuzio et al. 1996; Spandler et al. 2003; Martin et al. 2014). The formation of allanite due to lawsonite decomposition was proposed for natural HP rocks (e.g. Spandler et al. 2003) as well as demonstrated in experiments (e.g. Martin et al. 2014). However, the amount of LREE hosted in lawsonite may not be sufficient to form monazite in the Vestgötabreen eclogites. Thus, it is suggested that the additional source of LREE as well as P could have been provided by florencite,  $(REE)Al_3(PO_4)_2(OH)_6$ , an accessory mineral described from various HP rocks (Chopin and Monie 1984; Janots et al. 2006; Gabudianu Radulescu et al. 2009; Gautiez-Putallaz et al. 2016; Hoschek 2016). Barthélemy and Chopin (2006) performed experiments showing that florencite-(La) is stable up to 2.5 GPa at 500 °C and up to 1.5 GPa at 700 °C. Barnes et al. (2020a) reported remnants of florencite in the HP-LT phyllite from the Lower unit and interpreted it as a precursor for metamorphic monazite.

A tabular shape of pseudomorphs is an indicator for a former presence of lawsonite (Martin et al. 2014; Whitney et al. 2020) in rocks that are otherwise lacking remnants of

lawsonite. Based on the tabular shape of the allanite-epidote aggregates (Fig. 10a, b) and mineral composition, we suggest that monazite can be a product of both florencite and lawsonite break-down reactions. These two reactions could have happened at virtually similar P-T conditions between peak-P and peak-T stages (Fig. 13) at  $471 \pm 6$  Ma. This is the example of monazite growth linked to reactions between accessory minerals at high pressure conditions as an alternative mechanism to low grade alteration-driven growth (e.g. Schulze et al. 2015). Monazite in garnet-II could represent former inclusions of florencite, whereas the tabular and elongated pseudomorphs in the matrix may have been former lawsonite with inclusions of florencite. Once both minerals started breaking down, they were replaced by allanite-I (former lawsonite) and monazite (former florencite). Monazite started breaking down to allanite-II and apatite during decompression (Fig. 10c, e), which is a typical reaction described in metagranitoids and metapelites (e.g. Finger et al. 1998; Broska and Siman 1998; Majka and Budzyń 2006; Upadhyay and Pruseth 2012).

Summarizing our observations, we suggest that LREEbearing phases in the eclogite grew following the sequence florencite > monazite > allanite-II as proposed in previous work (e.g. Gabudianu Radulescu et al. 2009; Hoschek 2016). Additionally, decomposition of lawsonite may have contributed to the REE budget, as well as provided an additional source of Ca for garnet-II. We suggest that monazite growth at  $471 \pm 6$  Ma (Fig. 10h) happened between peak-P and peak-T conditions, essentially dating the peak P-T conditions for eclogite facies conditions in the Vestgötabreen Complex. In contrast, monazite in the blueschist forms rare inclusions captured only in the garnet core (Fig. 11). Its chemistry and textures are different from those in the eclogites. Although based on very few analyses, the monazite age of  $486 \pm 6$  Ma is consistent with the U-Pb zircon age and interpreted to record prograde growth before garnet nucleation, dating the subduction stage of the Vestgötabreen Complex.

An alternative scenario involves allanite and apatite as a reactants to form monazite. Allanite would be the stable phase during prograde metamorphism, and the growth of monazite could have happened between peak-P and peak-T conditions. Such reactions were described in HP rocks (e.g. Manzotti et al. 2018). Monazite decomposition to allanite and apatite coronas could have occurred during decompression (e.g. Finger et al. 1998; Broska and Siman 1998). Additionally, apatite as a precursor of monazite would explain the enrichment in MREE in monazite in eclogite compared to monazite in the blueschist.

Yet another possibility is that the growth of monazite in the eclogite is fully controlled by an influx of an external fluid. In such the case, this external fluid would have to be responsible for delivery of phosphorous and potential liberation of REE into the system. Although it is apparent that the garnet in in the eclogite must have interacted with a fluid, which is corroborated by its oscillatory zoning in the rim regions, it is hard to reconcile whether the fluid was external or not. A liberation of the sufficient amount of phosphorous and REE from dissolving garnet seems rather unlikely. If this was the case, then it would be expected to find metamorphic monazite in virtually every eclogitized blueschists worldwide. Hence, this scenario is not completely ruled out, but it is considered less plausible than the other possibilities presented above.

# Interpretation of the Lu–Hf garnet age

Interpreting the Lu-Hf age of the eclogite (KK14-55) is not straightforward. Major element distribution shows two garnet growth phases in all measured garnet grains (Fig. 5a, b), with a jump in composition from garnet-I to garnet-II. The trace element analyses (Fig. 5c) show a small Lu enrichment in the inner part of garnet-I (8 ppm) with decreasing content towards the outer part of garnet-I. It then rises again in the garnet-II with the highest contents of up to 35 ppm either in the very outer part of the rim or an oscillatory zoning within the rim. Five garnet separates were measured (466.4  $\pm$  4.5 Ma, MSWD = 9.4, n = 7, Fig. 12), of which only four fall onto a statistically good isochron with an omphacite separate and a whole rock. These four garnets give an isochron age of  $467.3 \pm 3.0$  Ma (MSWD = 3.3, n = 6; Fig. 12a). The other garnet separate gives a younger age of  $457.6 \pm 3.2$  Ma (MSWD = 0.59, n = 3; Fig. 12b). As the garnets seem to have two garnet growth generations, these ages most likely reflect mixed ages between garnet-I and garnet-II. With most of the Lu concentrated in the garnet-II, which are volumetrically the largest part of the garnet, the ages are probably biased towards garnet-II, with the garnet separate giving the young age being biased most strongly. The Lu-Hf results do not resolve the time of metamorphism, however zircon and monazite help to constrain the metamorphic evolution.

The blueschist sample (KK12-12E) gives an isochron age of  $471.1 \pm 3.8$  Ma (MSWD = 2.8, n = 10; Fig. 12c). The bellshaped Lu distribution (Fig. 7c) as well as the major element distribution (Fig. 7a, b) is typical for a garnet growth zonation and suggests garnets in this sample preserve their original growth zoning. The lower Lu rim profile shows smaller peaks towards the rims that are often observed in the growth zoning of garnet. Therefore, we interpret this age as dating garnet growth during blueschist-facies metamorphism.

# P-T-t evolutionary path of the Vestgötabreen Complex

Based on the P–T estimates and geochronology results we propose a new P–T–t path for the Vestgötabreen Upper unit

(Fig. 13). The zircon U–Pb age of  $482 \pm 10$  Ma is interpreted to represent Early Ordovician prograde metamorphism under blueschist-eclogite facies conditions (stage M1,  $1.6 \pm 0.3$  GPa at  $460 \pm 60$  °C). The peak pressure conditions, M2, reached  $2.3 \pm 0.3$  GPa and  $507 \pm 60$  °C, and were followed after  $467 \pm 3$  Ma by garnet-II growth in the eclogite at  $2.1 \pm 0.3$  GPa and  $553 \pm 60$  °C. The lawsonite and/or florencite break-down occurred between the peak-P (M2) and peak-T (M3) conditions around  $471 \pm 6$  Ma based on the monazite age from the eclogite. This interpretation is in agreement with textural observations as well as experimental data of Barthélemy and Chopin (2006). The monaziteallanite transition occurred later (post 471 Ma) during the decompression, but the exact timing is difficult to define because the allanite date has relatively high uncertainty  $(392 \pm 50 \text{ Ma}).$ 

Monazite from the blueschist is interpreted to form as a prograde phase during subduction at  $486 \pm 6$  Ma with subsequent peak-metamorphism up to blueschist facies  $(2.0\pm0.03$  GPa at  $500\pm30$  °C) conditions at  $471.1\pm3.8$  Ma (Lu–Hf dates, Figs. 12c, 13). Since blueschist and eclogite form lenses within the subduction mélange, the discrepancy between their P–T conditions could indicate that the blocks reached different depths. Lenses of blueschist and eclogite from the Upper unit likely were exhumed during the Middle Ordovician and assembled with the Lower unit rocks at  $454\pm6$  Ma (Barnes et al. 2020a).

# **Tectonic implications**

Based on the P-T-t history presented herein and in previous work (Hirajima et al. 1988; Agard et al. 2005), it is apparent that the Vestgötabreen Complex provides a unique record of a modern style cold subduction environment developed already in the earliest Paleozoic. This makes the studied HP-LT rocks a pivotal witness of deep elemental cycling and fluid migration potentially triggering intermediate depth earthquakes within the Iapetus Ocean realm, mostly thanks to the extraordinary level of preservation of the Vestgötabreen blueschists and eclogites. A similar style of HP-LT metamorphism, though not as well-preserved, has been reported from the Tsäkkok Lens of the Middle Allochthon in the Scandinavian Caledonides (e.g. Bukała et al. 2020). There, eclogitized blueschist facies rocks also yielded Early Ordovician ages of peak metamorphism (Barnes et al. 2020b; Fassmer et al. 2021; Root and Corfu 2012). These rocks reveal features characteristic for subduction within the lawsonite stability field, its breakdown and associated fluid infiltration leading to potentially seismicity-triggering hydrofracturing (Bukała et al. 2020). Accordingly, Barnes et al. (2020a) proposed that the HP-LT rocks of the Tsäkkok Lens in the Scandinavian Caledonides and the Vestgötabreen Complex in the Svalbard may have formed in the same subduction system. Majka et al. (2021) have investigated the Early Ordovician Thores Arc in the Pearva Terrane of northern Ellesmere Island, an island arc developed also in the northern Iapetus Ocean. These authors revealed that some plutonic rocks with a clear island arc affinity show inheritance of continental crustal material and proposed possible peri-Baltican ancestry for a continental sliver on which the intra-oceanic arc was built. Based on this and other lines of evidence, Majka et al. (2021) presented a tectonic model suggesting that the Virisen arc of the Scandinavian Caledonides (in their model the upper plate to the Tsäkkok eclogitized blueschist) and the Thores arc of the Pearya terrane (upper plate to the Vestgötabreen Complex) have indeed formed a continuous coherent subduction system proximal to the Baltica margin. The inheritance in zircon presented in this study is consistent with this idea because of general similarity of the obtained ages to those from the Thores Arc (Majka et al. 2021). This, in turn, suggests that at least some of the Vestgötabreen lithologies could have been introduced to the subducting plate via tectonic erosion mechanism. Notwithstanding the latter, it becomes evident that if the Vestgötabreen Complex was a part of the same subduction system as the Virisen arc (including the Tsäkkok blueschists), it could have been proximal to the Baltica margin, which breaks existing paradigms and adds another argument in a dispute favouring mixed Baltican and Laurentian provenance of terranes within the Southwestern Basement province of Svalbard (Wala et al. 2021). In addition, the location of this important HP-LT terrane in western Svalbard, with its probable Baltican affinity, supports the existence of a Silurian-Devonian strike-slip corridor developed during the waning stage of the Caledonian orogeny (e.g. Colpron and Nelson 2009; McClelland et al. 2022) and its major role in arrangement of crustal blocks in the High Arctic.

# Conclusions

The petrochronological study using garnet, zircon, monazite and allanite geochronology linked with conventional geothermobarometry, quartz-in-garnet elastic thermobarometry and Zr-in-rutile thermometry allows reconstruction of a pressure–temperature-time path for the high pressure-low temperature rocks of the Vestgötabreen Complex, Svalbard. Eclogite experienced a clockwise pressure–temperature path from prograde conditions at  $1.6 \pm 0.3$  GPa at  $460 \pm 60$  °C, through peak-pressure,  $2.3 \pm 0.3$  GPa and  $507 \pm 60$  °C, and to peak-temperature at  $2.1 \pm 0.3$  GPa and  $553 \pm 60$  °C. U–Pb zircon geochronology provides a date on the prograde part of the pressure–temperature path at  $482 \pm 10$  Ma. Monazite crystallized at  $471 \pm 6$  Ma as a response to the decomposition of accessory phases that happened between peak-pressure and peak-temperature conditions. Blueschist provides an indicative prograde monazite age of  $486 \pm 6$  Ma and Lu–Hf garnet peak-metamorphic age of  $471.1 \pm 4$  Ma. Peak-pressure metamorphic conditions are  $2.0 \pm 0.03$  GPa at  $500 \pm 30$  °C, defined by garnet. Moreover, it has been demonstrated that monazite can grow in mafic high-pressure rocks due to the breakdown of LREE-bearing minerals, florencite and/or lawsonite, as an alternative mechanism to metasomatic growth.

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**Data availability** All data are included in this article, its supplementary files and in the data repository.

#### Declarations

**Conflcit of interest** The authors declare that there are no competing interests.

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