Deciphering the tectonometamorphic history of subducted metapelites using quartz-in-garnet and Ti-in-quartz (QuiG–TiQ) geothermobarometry—A key for understanding burial in the Scandinavian Caledonides

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Abstract

The Seve Nappe Complex is a subduction-related high-grade metamorphic unit that was emplaced onto the margin of Baltica during Caledonian orogenesis. In this paper, the tectonometamorphic evolution of the Lower Seve Nappe in the Scandinavian Caledonides was characterized with the help of the continuous Collisional Orogeny in the Scandinavian Caledonides (COSC-1) drill core, using a combination of various P–T estimation techniques based on garnet–quartz mineral pairs (quartz-in-garnet and Ti-in-quartz [QuiG–TiQ]), conventional thermobarometry and thermodynamic modelling of phase equilibria. This multi-method approach yields complementary results and delivers critical data to constrain a comprehensive pressure–temperature–deformation–time (P–T–D–t) evolutionary path for the metasedimentary rocks of the Lower Seve Nappe. In the garnetiferous metasedimentary rocks, quartz inclusions in garnet preserve the P–T conditions of three distinct garnet growth stages corresponding to three metamorphic stages Ms1 to Ms3, including prograde and peak metamorphic conditions. Ms1 and Ms2 stages were constrained via quartz inclusions in garnet core and mantle. They are relatively close in the P–T space and could be considered as one single continuous prograde event occurring at epidote–amphibolite facies conditions of 460–520°C and 0.6–0.85 GPa. The growth of the garnet outermost rim defines the Ms3 stage at amphibolite facies conditions of 590–610°C and 1.13–1.18 GPa and corresponds to the peak metamorphic conditions. The microstructural analysis shows that the finite ductile strain pattern of the Lower Seve Nappe results from the superposition of four deformation phases. The initial phase D1 is defined by the S1 foliation that is still preserved as a curved inclusion trail in
the garnet core. The D2 phase initiated contemporaneously with garnet core growth and the development of muscovite–biotite–plagioclase S2 foliation. Garnet outermost rim growth marks the end of the prograde path and peak metamorphic conditions. This stage is overprinted by the D3 phase and Ms4 stage associated with the development of the main regional metamorphic and mylonitic fabric S3 associated with C–type shear bands along the retrograde path. Ms4 stage, which was constrained using traditional thermobarometric techniques, corresponds to the chemical re-equilibration of the metasedimentary minerals and occurred under amphibolite facies conditions at $\sim$570–610°C and 0.78–1.00 GPa. The D3 phase is then generally weakly to strongly overprinted by later lower grade deformation D4 phase at greenschist facies conditions (Ms5). $^{40}$Ar/$^{39}$Ar ages of syn-kinematic white mica and biotite indicate that the final stage of the thrusting of the Lower Seve Nappe and thus the timing of its emplacement onto the Offerdal Nappe occurred at c. 423 Ma. Collectively, these results are consistent with previous estimates of the timing and conditions of metamorphism derived from the Lower Seve Nappe especially in west-central Jämtland. However, application of QuiG–TiQ thermobarometry demonstrated that quartz inclusions in garnet can preserve different aspects of garnet growth, which are not accessible by traditional methods especially in complex terranes, and therefore provided new significant insights into the Lower Seve prograde evolution.

**KEYWORDS**

$^{40}$Ar/$^{39}$Ar dating, P–T–D–t evolution, quartz-in-garnet, Scandinavian Caledonides, Ti-in-quartz thermobarometry

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## 1 | INTRODUCTION

Metamorphic rocks recovered from fossil subduction zones provide key information on the processes occurring during the early stages of convergence associated with mountain building. They are excellent tracers of thermal and mechanical evolution that the lithosphere has experienced, as their mineralogical composition can record one or more stages of the Earth’s lithosphere thermal evolution. Deciphering the tectonometamorphic evolution of the rocks derived from exhumed metamorphic terranes along convergence zones may thus contribute to a better understanding of the crustal evolution and modelling of the major geodynamic processes occurring in these terranes. However, in most cases, the burial histories of subduction zone rocks are lost due to subsequent overprint at near-peak conditions or retrograde re-equilibration during exhumation. Therefore, the early prograde metamorphic evolution is often poorly constrained. Nevertheless, traces of early metamorphic stages can be preserved thanks to inclusions encapsulated in rigid mineral porphyroblasts.

Garnet represents a suitable example as such host for inclusions due to its broad stability field, resistance to deformation and largely chemically inert character with respect to the most common enclosed phase, which is quartz in continental crustal rocks. Therefore, quartz inclusions in garnet have significant potential as a tool for deciphering pressure–temperature (P–T) estimates of inclusion entrapment and thus garnet crystallization conditions independently from traditional geothermobarometric techniques based on major element exchange and net-transfer reactions. Pressure can be determined using quartz-in-garnet (QuiG) thermo-elastic models (Angel et al., 2015; Barkoff et al., 2017; Campomenosi et al., 2018; Gilio, Angel, & Alvaro, 2021; Gilio, Scambelluri, et al., 2021; Gonzalez et al., 2019; Kohn, 2014; Thomas & Spear, 2018), whereas temperature can be resolved based on the incorporation of Ti-in-quartz (TiQ), providing the basis for TiQ solubility models (Behr et al., 2011; Cruz-Uribe et al., 2017; Gonzalez et al., 2019; Grujic et al., 2011; Kidder et al., 2013; Korchinski et al., 2012; Nachlas et al., 2014; Şengün & Zack, 2016; Spear et al., 2012).
The Scandinavian Caledonides (Figure 1a), formed as a result of the collision of the continents of Laurentia and Baltica, is a good example of a convergent zone for which the knowledge of the early prograde evolution is poorly constrained. A prominent enigmatic issue remains the subduction of continental outer margins (Gee et al., 2020). This collisional orogeny is characterized by allochthonous thrust sheets that differ in composition, age of formation and grade of metamorphism, as they have experienced different metamorphic paths during...
their tectonic evolution. Over recent years, several petrological studies have demonstrated that one of those allochthonous sheets, the Seve Nappe Complex (SNC; part of the Middle Allochthon; Figure 1a,b), had experienced ultrahigh-pressure (UHP) metamorphism related to a deep subduction episode of the Baltic outer margin (e.g., Bukala et al., 2018; Fassmer et al., 2017; Janák et al., 2013; Klonowska et al., 2014, 2016, 2017; Majka et al., 2014; Petrik et al., 2019).

In the Åre region in west-central Jämtland (Figure 1c,d), the UHP metamorphism in paragneisses was documented by thermodynamic phase equilibrium modelling, mineral chemistry and the presence of microdiamond as inclusion in garnet in the Middle Seve Nappe (MSN) (Klonowska et al., 2017, 2021). However, in the Åre Synform, both the MSN and Lower Seve Nappe (LSN) are exposed (Figure 1d) and display an apparent inverted metamorphic gradient, where the base (LSN) consists of greenschist to amphibolite facies rocks (Giuntoli et al., 2018, 2020) in contrast to the (U)HP Middle Seve. Therefore, it is apparent that the different thrust sheets of the SNC have experienced different P–T conditions during the prograde to peak metamorphic history. This discrepancy could be either an effect of a burial of these two units of the Baltic passive margin to different depths in the same subduction system or indicate that the tectonometamorphic histories and provenance of the MSN and LSN may not be closely related to each other. To better understand the origin of the inverted metamorphic gradient and the emplacement process(es) of the lower and higher grade nappes of the SNC, it is critical to gain accurate knowledge of the geological processes that have affected those individual allochthonous units prior to their juxtaposition.

In 2014, the Collisional Orogeny in the Scandinavian Caledonides (COSC-1) scientific drilling project extracted a continuous drill core (~2.5 km deep) through the LSN (Lorenz et al., 2015; Figure 1c,d). In this study, we focus on pressure–temperature–deformation–time (P–T–D–t) evolutionary paths of the metamorphic rocks sampled by the deep drilling. In the garnetiferous metasedimentary rocks of the LSN, quartz inclusions in garnet have pre-erupted ultrahigh-pressure (UHP) metamorphism related to a deep subduction episode of the Baltic outer margin (e.g., Bukala et al., 2018; Fassmer et al., 2017; Janák et al., 2013; Klonowska et al., 2014, 2016, 2017; Majka et al., 2014; Petrik et al., 2019).

In addition, conventional geothermobarometry and thermodynamic modelling of phase equilibria provide constraints on the retrograde P–T path. In situ and single-grain fusion mica 40Ar/39Ar geochronology was conducted to constrain the timing of the exhumation of these rocks.

Thanks to the application of modern thermobarometric techniques based on garnet–quartz mineral pairs, it was possible to estimate garnet growth conditions using the same approach on multiple samples of different lithological varieties in spite of chemical modification and recrystallization during later events. Our study demonstrates that a combination of various P–T estimation techniques yields complementary results and delivers necessary data to constrain a comprehensive P–T–D–t evolutionary model for the metamorphosed rocks. More broadly, deciphering a comprehensive P–T–D–t history for the LSN brings new insights into the geodynamic model of the SNC especially in west-central Jämtland and contributes to a better understanding of the orogenic evolution of the Scandinavian Caledonides.

2 | GEOLOGICAL SETTING

The Scandinavian Caledonides (Figure 1a) formed as a result of late Cambrian to Ordovician-Silurian closure of the Iapetus Ocean and subsequent continental collision of the Balticate plate with the Laurentian plate in the late Silurian/Early Devonian (Gee et al., 2008). This continent–continent collisional event caused the amalgamation and thrusting of oceanic and continental allochthons (the Lower, Middle, Upper and Uppermost allochthons) up to 400 km towards the east-southeast over the autochthonous Baltica basement (Figure 1b; Gee et al., 2010, 2020; Rice & Anderson, 2016; Roberts, 2003; Roberts & Gee, 1985). The current architecture of the Caledonian nappes in central Sweden formed in the mid-crust by foreland-directed in-and out-of-sequence thrusting (Bender et al., 2018). The entire nappe pile was affected by pervasive, top-to-the-ESE general shear deformation at amphibolite to upper greenschist facies conditions (Bender et al., 2018; Bergman & Sjöström, 1997). The entire Scandinavian nappe pile is folded into SN-trending synforms and antiforms. This late stage of folding is related to either crustal extension and normal faulting or basement shortening occurring during the latest orogenic phases (Bergman & Sjöström, 1997; Rice & Anderson, 2016). Of these allochthons, thrust sheets of continental outer and outermost margin affinities (Middle Allochthon; Figure 1a,b), transported from locations west of the present Norwegian coast and thrust eastwards onto the Baltoscandian foreland basin and platform, are of higher metamorphic grade than underlying tectonic units and generally are more penetratively deformed (e.g., Gee et al., 2013, 2020). The Middle Allochthon includes (Figure 1; Gee et al., 2020): (1) the lower thrust sheets that comprise Palaeoproterozoic crystalline basement (e.g., the Tännäs Augen Gneiss Nappe) and greenschist facies Neo-proterozoic siliciclastic metasedimentary rocks (e.g., the Offerdal Nappe); (2) the Särv Nappe, a Cryogenian–Ediacaran succession intruded by c. 600 Ma dolerites with mid-ocean ridge basalt (MORB) composition (Hollocher et al., 2007); and, finally, (3) the upper sheets called the...
SNC, dominated by high-grade rocks (including HP and UHP units) comprising metasedimentary and metagneous rocks that represent the extended passive margin of Baltica (e.g., Andréasson, 1994; Gee et al., 2008; Roberts, 2003; Sjöström, 1983; Stephens, 1988). These rocks were subducted to mantle depths beneath a volcanic arc of Iapetus affinity during Caledonian orogenesis (e.g., Andréasson, 1987; Brueckner & van Roermund, 2004; Bukala et al., 2018, 2020; Gee, 1975; Janák et al., 2013; Klonowska et al., 2014, 2016, 2017; Klonska et al., 2021; Kullerud et al., 1990; Majka et al., 2014; Petrik et al., 2019; Santalier, 1988; Stephens & van Roermund, 1984).

The SNC can be further subdivided by the presence of internal thrust sheets into the LSN, MSN and Upper Seve Nappe (Sjöström et al., 1991; Trouw, 1973; Zachrisson & Sjöström, 1990). In the Jämtland region, several petrological studies have demonstrated HP to UHP metamorphism (Figure 1a) in the migmatitic paragneisses of the MSN by the presence of microdiamond (Klonowska et al., 2017, 2021; Majka et al., 2014; Petrik et al., 2019), and garnet peridotite and eclogite lenses in the LSN and MSN (Klonowska et al., 2016; Van Roermund, 1985). These rocks preserve mineral assemblages indicating peak P-T conditions of up to 4.1–4.2 GPa and 830–840°C (Gilio et al., 2015; Janák et al., 2013; Klonowska et al., 2017) at c. 470–460 Ma (Brueckner & Van Roermund, 2007; Fassmer et al., 2017). A more recent study (Walczak et al., 2022), utilizing a U–Pb age depth profiling of zircon grains, suggests that a UHP event, recorded in west-central Jämtland, occurred at 482.6 ± 3.8 Ma.

In the Åre region, the upper part of the nappe is mainly composed of carbonate–silicate rocks and marble, with numerous amphibolites and metadolerites, whereas the lower part is mainly composed of garnet-bearing mica schists and metapsammite (Figure 1c,d). The overall unit was intruded by felsic rocks, which were subsequently deformed and metamorphosed together with the host rock (Figures 2 and 3; Li et al., 2021). The MSN is in this region (Figure 1c,d), the structurally highest unit (Arn bom, 1980), and comprises a well-exposed klippe of garnetiferous paragneisses, locally with microdiamonds and migmatites (Klonowska et al., 2017, 2021). The migmatization indicates that the UHP rocks were exhumed to lower crustal levels at 442–436 Ma (Ladenberger et al., 2014; Majka et al., 2012; Walczak et al., 2022). Subsequently, all thrust sheets in the SNC were affected by middle pressure metamorphism at ~1.0–0.6 GPa and ~600°C (Arn bom, 1980; Bergman, 1992; Giuntoli et al., 2018, 2020) as the MSN was emplaced onto the LSN along major shear zones (Figure 1c,d). Several studies (Bender et al., 2019; Dallmeyer, 1990; Dallmeyer et al., 1985; Grimmer et al., 2015; Gromet et al., 1996; Hacker & Gans, 2005; Majka et al., 2012) have shown that the tectonic control between the MSN and LSN was active between 435 and 423 Ma. Successively, the LSN experienced further decompression to 0.8–0.5 GPa and 600–500°C at 417 ± 9 Ma (Giuntoli et al., 2020). Finally, late-stage km-scale open folds with NNE–SSW-trending axes parallel to the strike of the orogen (Sjöström, 1978; Trouw, 1973) overprinted the nappe stack. In central Jämtland, this late stage constitutes major map-scale features (Hedin et al., 2014; Palm et al., 1991) and controls the dipping direction of the regional foliation plane. In the Åre Synform (Figure 1c,d), two generations of foliation were identified: (1) a main foliation with a dip direction towards E-NE and mean dip value of 30° (Bender et al., 2018), related to upper greenschist to amphibolite facies metamorphism overprinted by (2) a lower grade foliation subparallel to it and related to lower to sub-greenschist facies metamorphism. Both foliations develop stretching lineations that indicate pervasive top-to-the-ESE general shear during retrograde metamorphism (Bender et al., 2018).

The COSC-1 borehole (Figure 3) is located in the gently W-dipping eastern limb of the Åre Synform (Lorenz et al., 2015; see location in Figure 1c). The position of the samples used in this study is shown in Figure 3 together with the samples previously studied by Giuntoli et al. (2018, 2020). The upper part of the drill core, from the surface down to ~1600 m, is characterized by alternating layers of deformed garnet orthogneiss, garnet calc-silicate, marble and amphibolite. Below 1600 m, garnet mica schist and metasandstone are the most prominent lithologies alternating with layers of deformed felsic gneiss and amphibolite (Figure 3). The rocks show strongly deformed mylonitic fabrics from ~1600 m to the bottom of the core at 2500 m depth (Giuntoli et al., 2018; Hedin et al., 2016; Kästner et al., 2021; Merz et al., 2019). The bottom of the core is interpreted as representing the basal shear zone juxtaposing the LSN with the Offerdal Nappe (Gee et al., 2020). The main regional foliation in the core is generally subhorizontal, locally affected by recumbent folds (Wenning et al., 2017) and shear bands associated with S/C/C′ fabrics (see core scans in Figure 3). These folds were coeval with top-to-the-ESE shearing as defined by field-based studies (Bender et al., 2018), with subhorizontal fold axial surfaces parallel to the main mylonitic foliation and to the nappe transport direction.

The Så quarry, located ~7 km southeast of Åre and 4 km south of COSC-1 drill site (see location in Figure 1c), provides an excellent additional exposure of the LSN (Li et al., 2021). This 20–30 m thick section through the central part of the LSN (Figure 2) permits a better understanding of the structural relationships between the deformed orthogneiss, the
amphibolite boudins and the calcsilicate host rock (Figure 2). The amphibolite and the orthogneiss appear as boudins that are highly deformed together with the host rock and display ESE-trending isoclinal folds. Both intrusive lithologies exhibit a clear foliation, which in most cases is concordant with the main regional foliation that is subhorizontal to moderately dipping towards the southeast and striking N30°. However, the orthogneissic foliation in Figure 2b seems to crosscut an early foliation present in the amphibolite boudins at a low angle. Moreover, this metaigneous rock can even occur locally as intrusion within the mafic boudin (Figure 2c). These features suggest that the orthogneiss must have intruded the metasedimentary rocks after the amphibolites but before the main regional foliation started to develop. Zircon U–Pb dating was reported recently from two samples of deformed felsic gneisses from the Så quarry (Li et al., 2021) and yielded consistent 468–470 Ma magmatic ages. These results are of particular importance for the tectonic history of the SNC because they provide clear evidence of a pre-Silurian magmatic event, previously unknown in the SNC.
FIGURE 3  Illustrative geological log of the Collisional Orogeny in the Scandinavian Caledonides (COSC-1) drill core and example photographs of the core samples. The numbers represent all of the samples pre-selected for this study; bold text highlights the samples chosen for the detailed investigations (see Tables 1 and S1). The COSC-1 samples previously studied in the interest of revealing the metamorphic history of the Lower Seve Nappe (Giuntoli et al., 2018, 2020) have been represented with green star for amphibolite samples and yellow stars for mica schist samples.
3 | SAMPLING, SAMPLE PREPARATION AND ANALYTICAL TECHNIQUES

3.1 | COSC-1 drill core and sampling

We selected 27 samples of garnet-bearing metamorphic rocks from the COSC-1 drill core between depths of 146 and 2491 m (Figure 3). The sampled lithologies include calc-silicates, amphibolite with garnet-bearing lenses of calc-silicates, orthogneisses and garnet-bearing mica schists. From petrographic observations, we choose a suitable set of samples (n = 11) for further geothermo-barometric and geochronological investigations. It includes mostly garnetiferous mica schists with and without amphibole from the lower part of the drill core (1575 to 2491 m), namely, samples 516-2, 582-2, 585-5, 595-6, 615-2, 629-1, 641-6, 683-4, 690-6, 691-2 and 695-5 (Tables 1 and S1). Transmitted light microscope was used to determine the presence and distribution of quartz inclusions in garnet for the QuiG–TiQ thermobarometric approach. SE images were used to show distribution at the thin-section scale and to determine relationships to structural fabric. Quartz inclusions were selected based on their size, ranging in diameter from 5 to 20 μm, in areas absent of cracks and other mineral phases. Thin sections (~30–40 μm) and thick sections (~80 μm) were prepared for TiQ and QuiG methods, respectively, from the mirrored parts of the samples. Inclusions selected for elastic thermobarometry were mostly spherical and located under the surface of the garnet >10 μm from interfaces (e.g., cracks and polished surface) to reduce the chance of measuring modified inclusions.

3.2 | Electron microprobe

Mineral composition, X-ray compositional maps of major elements in garnet and bulk rock chemistry were obtained using a Jeol JXA8530F Hyperprobe field emission electron probe microanalyser (FE-EPMA) at the Department of Earth Sciences, Uppsala University, Sweden.

Mineral composition and bulk rock chemistry were measured only for one mica schist 516-2 (Tables 2 and 3). The operating conditions of Jeol JXA8530F Hyperprobe for spot analyses were as follows: 10 nA beam current with 15 kV accelerating voltage and counting times of 10 s on peaks and 5 s on background positions, and 1 to 10 μm beam depending on the mineral analysed. Mineral standards (Si, Ca—wollastonite; Na—albite; K—orthoclase; Mn, Ti—pyrophanite; Fe—fayalite; Mg—pyrope; for garnet analyses) and pure element oxides (Al₂O₃, MgO, Cr₂O₃ and NiO) were used for calibration. All elements were analysed by Kα spectral lines. The bulk composition was collected using continuous energy dispersive spectrometer (EDS) scans of a large portion of the thin section.

EPMA X-ray compositional maps of calcium, magnesium and manganese were collected on selected garnets (samples 516-2, 615-2, 629-1, 683-4 and 690-6; Table 1) to identify and characterize the internal zoning patterns in each grain and therefore determine the location of quartz grains included in these garnets. Maps of aluminium, calcium and sodium were also collected on one selected grain of plagioclase in sample 615-2. For this purpose, the same FE-EPMA instrument as for the mineral composition was used and the mapping was performed under the following conditions: dwell time of 80–100 ms, beam current of 20–40 nA, accelerating voltage of 15 kV, beam diameter of 1–3 μm and one accumulation. The same mineral standards and spectral lines were used as for the spot analyses (see above).

Titanium concentrations in quartz (Table S2) were measured using a Cameca SXFive electron microprobe at Syracuse University, Syracuse, NY, USA. Measurements were performed with 15 kV accelerating voltage and 200 nA beam current using a fully focused electron beam. X-ray intensities of Ti Kα were acquired simultaneously on four spectrometers equipped with large PET diffraction crystals and aggregated to improve counting statistics. Counting times of 200 s on peak and 100 s on high and low background yielded a detection limit of 5 ppm Ti for individual measurements. A blank correction for TiQ following Donovan et al. (2011) was applied using quartz from Herkimer County, NY, which has been independently measured with secondary-ion mass spectrometry (SIMS) to contain 0.005 ppm Ti (Kidder et al., 2013). Accuracy of trace TiQ measurements was confirmed by analysis of in-house synthetic quartz standards in addition to the sample of smoky quartz of Audétat et al. (2015).

3.3 | Raman spectroscopy

Micro-Raman spectroscopy was performed on 10 samples (Table 1). Quartz inclusions in garnet were analysed using a confocal laser Raman microspectrometer, Horiba instrument LabRAM HR 800 (Horiba Jobin Yvon), equipped with a multichannel air-cooled (−70°C) 1024 × 256 pixel CCD (charge-coupled device) detector, at the Department of Geological Sciences, Stockholm.
<table>
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<th>Sample name (in the text)</th>
<th>Top depth (m)</th>
<th>Bottom depth (m)</th>
<th>Lithology</th>
<th>Mineral paragenesis</th>
<th>Grt X-ray map</th>
<th>Pl X-ray map</th>
<th>QuiG</th>
<th>TiQ</th>
<th>GBMP</th>
<th>Pseudo modelling</th>
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Note: The grey shading was used to indicate/specify which type of analysis has been performed on every samples.
Abbreviations: GBMP, garnet-biotite-muscovite-plagioclase; QuiG, quartz-in-garnet; TiQ, Ti-in-quartz.
| Mineral Zone/position | Garnet | | | | | | Feldspar |
|----------------------|--------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|
|                      | Core   | Mantle         | Outer rim      | P1 I (BSE dark) | P1 I (BSE light) | P1 II | Kfs |
| SiO₂                 | 36.54  | 36.31  | 36.62 | 36.77 | 37.22 | 36.89 | 37.30 | 68.47 | 68.28 | 66.48 | 65.79 | 63.94 | 65.61 | 64.14 |
| TiO₂                 | 0.15   | 0.21   | 0.19  | 0.11  | 0.15  | 0.10  | 0.07  | 0.19  | 0.02  | 0.00  | 0.02  | 0.00  | 0.04  | 0.00  | 0.00  |
| Cr₂O₃                | 0.00   | 0.01   | 0.00  | 0.02  | 0.00  | 0.08  | 0.04  | 0.00  | -     | -     | -     | -     | -     | -     | -     |
| FeO                  | 20.37  | 20.68  | 20.59 | 21.45 | 21.28 | 21.49 | 21.46 | 21.61 | 21.78 | -     | -     | -     | -     | -     | -     |
| MnO                  | 14.44  | 13.96  | 14.21 | 13.53 | 13.19 | 12.89 | 12.49 | 12.47 | -     | -     | -     | -     | -     | -     | -     |
| MgO                  | 1.63   | 1.65   | 1.53  | 1.52  | 1.62  | 1.54  | 1.52  | 1.51  | 1.50  | -     | -     | -     | -     | -     | -     |
| CaO                  | 5.42   | 5.62   | 6.18  | 6.47  | 6.72  | 6.95  | 7.22  | 6.72  | 6.67  | 1.39  | 1.62  | 2.42  | 2.8  | 3.61  | 0.06  |
| Na₂O                 | 0.18   | 0.15   | 0.10  | 0.06  | 0.02  | 0.00  | 0.03  | 0.07  | 9.68  | 10.19 | 9.59  | 9.37  | 9.37  | 9.79  | 0.07  |
| K₂O                  | 0.00   | 0.00   | 0.01  | 0.03  | 0.02  | 0.00  | 0.01  | 0.04  | 0.05  | 0.04  | 0.07  | 0.10  | 0.12  | 0.10  | 16.08 |
| Total                | 99.50  | 99.33  | 100.13 | 100.85 | 100.82 | 100.87 | 100.65 | 100.52 | 100.39 | 100.68 | 101.18 | 99.97 | 99.62 | 100.68 | 98.68 |

Note: Alm = 100*Fe/(Fe + Mg + Ca + Mn); Sps = 100*Mn/(Fe + Mg + Ca + Mn); Prp = 100*Mg/(Fe + Mg + Ca + Mn); Grs = 100*Ca/(Fe + Mg + Ca + Mn); X Fe = Fe/(Fe + Mg).

Abbreviation: BSE, back-scattered electron.
University. The instrument has a spectral resolution of \(0.3 \text{ cm}^{-1}\). The holographic grating was set to 1800 grooves/mm. Excitation was provided by an Ar-ion laser \((\lambda = 514 \text{ nm})\) source. Spectra were recorded using a laser power of 1 mW at the sample surface. Sampling was carried out using an Olympus BX41 microscope coupled to the instrument and the laser beam was focused through a 100× objective to obtain a spot size of 1 \(\mu\text{m}\). Collection times for the Raman spectra are 10 accumulations of 10 s each. The accuracy of the instrument was controlled by the repeated use of a silicon (Si) reference with a Raman line at 520.7 \(\text{cm}^{-1}\). The Raman spectra were recorded with LabSpec 5 software and subsequently baseline-corrected for the continuum luminescence background when necessary, temperature-reduced to account for the Bose–Einstein occupation factor (Kuzmany, 2009) and normalized to the acquisition time. Peak positions, full-widths at half maximum (FWHMs) and integrated intensities were determined from fits with pseudo-Voigt functions. For each of the selected Raman bands, we determined the change in peak position \(\Delta \omega\) as the difference between the Raman shift of the inclusion \(\omega_i\) from that of an unstrained reference crystal \(\omega_0\). As standards, we used a free (unstrained) quartz crystal, measured multiple times during each measurement session at ambient pressure and temperature (0.1 MPa and 20°C) to eliminate shifts in peak positions due to instrumental drift and/or minor changes in room temperature. The \(\omega_0\) values were averaged and then subtracted from the \(\omega_i\) of the strained inclusions analysed in between two consecutive standard measurements. The \(\Delta \omega\) values of each mode of quartz inclusions and unstrained standards are listed in Table S3. Finally, sets of \(\Delta \omega\) for each inclusion (modes \(\omega_{128}, \omega_{206}, \omega_{265}\) and \(\omega_{464}\), as they are generally unaffected

### Table 3

Representative microprobe analyses of white mica and biotite from sample 516-2

<table>
<thead>
<tr>
<th>Mineral Zone/position</th>
<th>Wm I incl. in Pl (BSE dark)</th>
<th>Wm II</th>
<th>Wm III</th>
<th>Bt I</th>
<th>Bt II</th>
<th>Bt III</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>48.82 49.37 49.78 47.16</td>
<td>46.64</td>
<td>47.05</td>
<td>45.79</td>
<td>37.38</td>
<td>36.38</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.78  0.35 0.43 0.71</td>
<td>0.59</td>
<td>0.67</td>
<td>0.50</td>
<td>1.93</td>
<td>1.99</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>28.30 27.20 27.20 30.82</td>
<td>31.44</td>
<td>30.72 30.87</td>
<td>17.01</td>
<td>17.22</td>
<td>17.12</td>
</tr>
<tr>
<td>FeO</td>
<td>3.21  3.40 03.06 3.86</td>
<td>3.40</td>
<td>3.32 3.59</td>
<td>18.00</td>
<td>18.06</td>
<td>18.17</td>
</tr>
<tr>
<td>MnO</td>
<td>0.00  0.00 0.04 0.12</td>
<td>0.04</td>
<td>0.02 0.09</td>
<td>0.34</td>
<td>0.30</td>
<td>0.23</td>
</tr>
<tr>
<td>MgO</td>
<td>2.33  2.73 2.75 1.60</td>
<td>1.36</td>
<td>1.39 1.54</td>
<td>11.04</td>
<td>11.11</td>
<td>11.4</td>
</tr>
<tr>
<td>CaO</td>
<td>0.01  0.00 0.00 0.01</td>
<td>0.00</td>
<td>0.01 0.00</td>
<td>0.02</td>
<td>0.00</td>
<td>0.01</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.62  0.60 0.48 0.92</td>
<td>0.86</td>
<td>0.67 0.81</td>
<td>0.29</td>
<td>0.23</td>
<td>0.18</td>
</tr>
<tr>
<td>F</td>
<td>0.20  0.31 0.36 - -</td>
<td>- -</td>
<td>- -   - -</td>
<td>0.26</td>
<td>0.25</td>
<td>- -</td>
</tr>
</tbody>
</table>

**Total** | 94.43 | 94.02 | 94.04 | 94.76 | 93.52 | 93.76 | 93.09 | 95.55 | 94.88 | 95.72 | 95.11 | 94.77 | 94.01

**O** | 11.00 | 11.00 | 11.00 | 11.00 | 11.00 | 11.00 | 11.00 | 11.00 | 11.00 | 11.00 | 11.00 | 11.00 | 11.00

**Si** | 3.31 | 3.37 | 3.38 | 3.19 | 3.18 | 3.21 | 3.16 | 2.82 | 2.77 | 2.84 | 2.77 | 2.76 | 2.77

**Ti** | 0.04 | 0.02 | 0.02 | 0.04 | 0.03 | 0.03 | 0.03 | 0.11 | 0.11 | 0.11 | 0.11 | 0.09 | 0.09

**Al** | 2.26 | 2.19 | 2.18 | 2.46 | 2.53 | 2.47 | 2.51 | 1.51 | 1.55 | 1.51 | 1.54 | 1.49 | 1.51

**Fe²⁺** | 0.18 | 0.19 | 0.17 | 0.22 | 0.19 | 0.19 | 0.21 | 1.14 | 1.15 | 1.14 | 1.15 | 1.21 | 1.2

**Mn** | 0.00 | 0.00 | 0.00 | 0.01 | 0.00 | 0.00 | 0.01 | 0.02 | 0.02 | 0.01 | 0.02 | 0.02 | 0.02

**Mg** | 0.24 | 0.28 | 0.28 | 0.16 | 0.14 | 0.14 | 0.16 | 1.24 | 1.26 | 1.23 | 1.29 | 1.35 | 1.33

**Ca** | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00

**Na** | 0.08 | 0.08 | 0.06 | 0.12 | 0.11 | 0.09 | 0.11 | 0.04 | 0.03 | 0.03 | 0.03 | 0.03 | 0.03

**K** | 0.88 | 0.87 | 0.86 | 0.82 | 0.80 | 0.86 | 0.87 | 0.89 | 0.91 | 0.87 | 0.91 | 0.93 | 0.93

**F** | 0.04 | 0.07 | 0.08 | - - | - - | - - | - - | 0.06 | 0.06 | - - | 0.00 | 0.00 | 0.00

**Total** | 7.00 | 7.00 | 6.97 | 07.02 | 6.98 | 7.00 | 7.05 | 7.78 | 7.81 | 7.74 | 7.82 | 7.88 | 7.87

**X_{Mg}** | 0.56 | 0.59 | 0.62 | 0.43 | 0.42 | 0.43 | 0.43 | 0.52 | 0.52 | 0.52 | 0.53 | 0.53 | 0.53

*Note: \(X_{Mg} = Mg/(Fe + Mg)\).*

*Abbreviation: BSE, back-scattered electron.*
by overlap with modes of the host garnet) have been used to determine strain using the software stRAinMAN (Angel et al., 2019) by employing the Grüneisen tensor for quartz (Murri et al., 2018). The strain of each inclusion was converted to stress using the elastic tensors at room PT for quartz (Wang et al., 2015). The $P_{\text{inc}}$ was calculated from the stress tensor as the negative of the mean normal stress ($P_{\text{inc}} = -(\sigma_1 + \sigma_2 + \sigma_3)/3$). The uncertainty on the $P_{\text{inc}}$ was calculated from the variance–covariance matrix of the strain using the method described in Mazzucchelli et al. (2021).

### 3.4 In situ and single-grain fusion $^{40}\text{Ar}/^{39}\text{Ar}$ analytical methods

Two samples for in situ intragrain laser ablation (samples 595-6 and 695-5) were prepared by adhesion of rock chips containing muscovite and biotite to a glass plate, followed by grinding and polishing to a thickness of ~500 μm and then cutting into circular sections of ~1 cm in diameter. The rock chips were then cleaned with acetone before being removed from the glass. For single-grain fusion analysis, inclusion-free, unaltered muscovites of 60–200 μm were obtained by hand crushing, washing and sieving, followed by hand-picking under a binocular microscope. Mineral separates were irradiated in the CLICIT facility of the Oregon State TRIGA Reactor, USA. Standards and unknowns were placed in 2 mm deep wells in 18 mm diameter aluminum disks, with standards placed strategically so that the lateral neutron flux gradients across the disk could be evaluated. Fish Canyon sanidine (28.2 Ma; Kuiper et al., 2008) and GA1550 biotite (98.5 Ma; Spell & McDougall, 2003) were used as standards. Minerals were baked with an infrared lamp for 48 h for in situ samples and 24 h for single-grain samples.

The in situ intragrain laser ablation and single-grain fusion $^{40}\text{Ar}/^{39}\text{Ar}$ analyses were performed at the University of Manitoba using a multi-collector Thermo Fisher Scientific ARGUS VI mass spectrometer that is linked to a stainless steel Thermo Fisher Scientific extraction/purification line, Photon Machines (Analyte Excite) 193 nm laser (for in situ analyses) and Photon Machines (55 W) Fusions 10.6 CO2 laser (for single-grain fusion analyses). Reactive gases were removed by three NP-10 SAES getters prior to being admitted to the mass spectrometer. Argon isotopes of mass 40 to 37 were measured using Faraday detectors with low noise $1 \times 10^{12}$ Ω resistors and that of mass 36 was measured using a compact discrete dynode detector. The sensitivity for argon measurements is ~3.612 × 10^{-17} mol/FA as determined from measured aliquots of Fish Canyon sanidine (e.g., Dazé et al., 2003; Kuiper et al., 2008). The five argon isotopes were measured simultaneously over a period of 6 min. Mass discriminations were monitored by online analysis of air pipettes. A value of 295.5 was used for the atmospheric $^{40}\text{Ar}/^{36}\text{Ar}$ ratio (Steiger & Jäger, 1977). Corrections are made for neutron-induced $^{40}\text{Ar}$ from potassium, $^{39}\text{Ar}$ and $^{36}\text{Ar}$ from calcium and $^{36}\text{Ar}$ from chlorine. The decay constants used were those recommended by Steiger and Jäger (1977). A raster size of ~100 μm × 100 μm with a laser spot size of 50 μm in diameter was used for in situ ablation of the samples. The results of the dating are given in Table 5 and the complete isotopic data are provided in Table S4.

### 4 THERMOBAROMETRY

#### 4.1 TiQ and QuiG thermobarometry

Combined TiQ (Thomas et al., 2010) and QuiG (Angel et al., 2017; Kohn, 2014) Raman-shift thermobarometry was used to estimate the temperature and pressure of garnet growth. A similar methodology was previously employed by Gonzalez et al. (2019) and Johnson et al. (2021).

TiQ thermobarometry, applicable to geological conditions up to ~2.0 GPa (Thomas et al., 2010, 2015), uses Ti-Si substitution in quartz to estimate the temperature at which quartz crystallized. Therefore, it requires the in situ measurement of Ti concentrations in quartz. Titanium mass fractions in quartz from high-pressure low-temperature (HP-LT) terranes, found in many subduction zone lithologies, are typically measured by SIMS (Ashley et al., 2013; Behr et al., 2011; Kidder et al., 2013; Thomas et al., 2010), laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) (e.g., Cruz-Uribe et al., 2017) or EMPA (Donovan et al., 2011). We used an EMPA instrument and methodology to minimize detection limits and enable analysis of low level Ti concentrations in quartz from HP-LT terranes.

TiQ temperatures reported in this paper were calculated at different pressures using the following solubility equation of Thomas et al. (2010):

$$
T(°C)_{Qz} = \frac{(60,952 \pm 3177) + (1741 \pm 63)P}{(1.52 \pm 0.03) - R \ln (X_{TIO_2}^{Qz}) + RL_{aTIO_2}} - 273.15
$$

where $R$ is the gas constant 8.3145 J/K, $T$ is temperature in Celsius, $X_{TIO_2}^{Qz}$ is the mole fraction of TiO$_2$ in quartz and $a$TiO$_2$ is the activity of TiO$_2$ in the system.
This calibration was preferred over others (e.g., Huang & Audétat, 2012; Wark & Watson, 2006) because the equation (1) includes the effect of both pressure and temperature on the solubility of TiO$_2$ and (2) considers variable activity of TiO$_2$ for quartz crystallization. Ashley and Law (2015) demonstrated that for pelitic compositions, although large portions of P–T space have titanias activities $\sim$1.0, some portions where rutile (or ilmenite) is absent or occurs in very low abundance (200°C $< T < 500°C$ and 0.2 GPa $< P < 0.7$ GPa) are substantially lower than this. As a consequence, large errors occur in the projection of Ti isopleths relative to an assumed fixed activity.

Quartz inclusions in garnet are also suitable for barometry. Because of the difference in the elastic properties of these two minerals, quartz completely entrapped inside garnet will develop a residual stress (Rosenfeld & Chase, 1961), from which the entrapment pressure can be calculated. Raman spectroscopic measurements of quartz inclusions were used to calculate residual stress of entrapped inclusions from Raman band shifts of quartz.

A mineral inclusion can develop residual pressures upon exhumation due to the difference of its thermodynamic properties from those of the surrounding mineral host (Angel et al., 2014; Rosenfeld & Chase, 1961). Consider a soft inclusion in a stiffer host (e.g., QuiG) entrapped at a certain PT$_{HP}$ condition and exhumed to the surface (PT$_0$). At PT$_0$ conditions, both host and inclusion have a larger volume than at entrapment, due to pressure being released upon exhumation ($\Delta V > 0$). However, the volume increase of the softer quartz should be greater than that of the stiffer host ($\Delta V_{qz} > \Delta V_{grt}$). The host garnet constrains the quartz inclusion into a smaller volume than a free quartz crystal, straining and pressurizing the inclusion. Knowing the inclusion pressure ($P_{inc}$) at room conditions and the thermodynamic properties of host and inclusion allows one to backcalculate a line in PT space of possible entrapment conditions: the isomeke (Angel et al., 2017, 2014; Rosenfeld & Chase, 1961; Tajčmanová et al., 2021).

### 4.2 Other thermobarometric methods

Additional estimates of metamorphic P–T conditions were calculated for one preselected mica schist sample (516-2) using a combination of phase equilibria modelling (P–T pseudosection) and major element thermobarometry (garnet–biotite–muscovite–plagioclase [GBMP] net-transfer barometer; Wu, 2015).

Within this sample, the mineral association Grt + Bt + Ms + Pl was commonly observed to be in textural equilibrium, indicating that the GBMP geobarometer can be used to estimate the P–T conditions of final

<table>
<thead>
<tr>
<th>Table 4</th>
<th>Summary of the parameters used for the thermodynamic modelling for sample 516-2</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Thermodynamic component list</strong></td>
<td><strong>Bulk rock composition (wt%)</strong></td>
</tr>
<tr>
<td>MnO</td>
<td>0.22</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>6.32</td>
</tr>
<tr>
<td>CaO</td>
<td>2.27</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>3.75</td>
</tr>
<tr>
<td>FeO</td>
<td>5.19</td>
</tr>
<tr>
<td>MgO</td>
<td>3.00</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>20.27</td>
</tr>
<tr>
<td>SiO$_2$</td>
<td>58.16</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>0.82</td>
</tr>
<tr>
<td>O$_2$</td>
<td>0.15</td>
</tr>
<tr>
<td>H$_2$O</td>
<td>Saturated</td>
</tr>
</tbody>
</table>

Note: Abbreviations with the lower case are used for pure phases and upper case for solid solutions.

garnet-matrix equilibrium. Multiple mineral pairs have been selected based on their structural position and textural relationships. A P-T pseudosection was computed in the MnO–Na2O–CaO–K2O–FeO–MgO–Al2O3–SiO2–H2O–TiO2–O2 (MnNCKFMASHTO) system, under water-saturated conditions, using the thermodynamic data of Holland and Powell (2011) (hp11ver.dat) and the Perple_X software (Version 6.9.1; Connolly, 2005). The content of free O2 was set to 0.15 wt% (Table S3). The following solid solution models were closely predicted phases of interest and their proportions (Figure S3). The following solid solution models were used: garnet, chlorite, white mica, biotite (White et al., 2014; biotite updated by Holland et al., 2018), ilmenite (White et al., 2007), plagioclase, K-feldspar (Holland & Powell, 2003), amphibole (Green et al., 2016), epidote (Holland & Powell, 2011) and clinopyroxene (Green et al., 2007).

5 | RESULTS

5.1 | Microstructures and petrography

The following petrographic and microstructural description is based on 11 mica schist samples initially selected for geothermobarometric and geochronological investigations (Table 1). The microstructural observations demonstrate that the finite ductile strain pattern of the LSN results from the superposition of four deformation phases (D1 to D4) occurring at different metamorphic conditions (Ms1 to Ms5) similar to those described by Sjöström (1983) for the LSN in the nearby Handöl-Storlien area and by Merz et al. (2019) and Giuntoli et al. (2020) in studies of the COSC-1 core. In the lower part of the drill core, from 1575 to 2491 m, sample spacing (90 m on average) illustrates a relatively detailed picture of the strain gradient. It increases towards the bottom of the drill core, which corresponds to the tectonic contact that juxtaposed the Seve Nappe with the Offerdal Nappe. The deformation is heterogenous and partitioned between low-strain lenses and high-strain mylonitic zones. Depending on the protolith composition (± abundance of mica) and intensity of strain, various fabrics and kinematic indicators can be observed along the drill core (Figures 3 and 4).

5.1.1 | D1 deformation phase

The D1 deformation is associated with the development of a S1 foliation (Figure 5a–d), which is preserved as straight, fine-grained inclusion trails in plagioclase porphyroblasts (sample 516-2; Figure 5a) and curved inclusion trails in the core of garnet porphyroblasts (all samples, shown in Figure 5c for sample 516-2) and sometimes in the core of amphibole porphyroblasts (samples 582-2 and 595-6; Figure 5d). These inclusion patterns suggest that the plagioclase growth was syn-S1 to intertectonic and occurred between the D1 and the D2 deformation phases whereas the garnet and amphibole core growth is syntectonic and probably occurred during D2. Plagioclase porphyroblasts contain variable amounts and types of inclusions, mainly quartz, biotite and sulphides/opaque minerals but white mica, ilmenite, rutile, zircon, allanite, allanite–epidote, chlorite, monazite and Ca–Mg–Fe–Mn carbonates can also be observed (samples 641-6 and 683-4). Garnet porphyroblasts contain mainly quartz, rutile, ilmenite/other opaque minerals but also rare biotite, white mica, plagioclase (samples 585-5 and 615-2), allanite–epidote, allanite and calcite (sample 615-2).

5.1.2 | D2 deformation phase

D1-related fabrics are overprinted by the D2 deformation phase, which, in the early stage, is characterized by the development of a S2 crenulation cleavage and F2 folds (Figures 4 and 5b,c). Some samples (e.g., 516-2) preserve S1 folded by F2 and display a weak development of S2 parallel to the axial surfaces of F2 folds (Figure 5b). The latter, which are rarely preserved, are synchronous with the growth of the garnet and amphibole cores as explained above. The growth of garnet rims outlines the final stage of the D2 deformation when S1 is transposed into S2 (Figure 5c). S2 is predominantly defined by polymineralic layers of white mica-fish, biotite and plagioclase ± amphibole (Figure 5e) within which abundant porphyroblasts of (1) syn-S1 to intertectonic plagioclase (500–5000 μm length) and (2) syntectonic garnet (250–2500 μm) with early D2 cores and late D2 rims exist (Figure 5). In sample 595-6, garnet porphyroblasts form clusters (Figure 4). In a few strongly deformed samples, porphyroblasts of (3) syntectonic hornblende (samples 582-2 and 595-6) occur as amphibole fish with monoclinoic shape oriented parallel to the S2 foliation (up to ~1200 μm) or as larger rotated porphyroblasts (~4000 μm) with curved inclusion trails wrapped by the S2 foliation (Figure 5d). In samples 582-2 and 629-1, a second generation of post-D2 smaller euhedral garnets (60–700 μm) can be observed (Figure 5f). This garnet is strongly chemically zoned (see Section 5.2.2 and Figure 6f) and mainly inclusion free (sample 582-2) but occasionally contain inclusions of quartz, rutile, ilmenite, zoned epidote and white mica. Those inclusions do not

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define any internal inclusion fabric \( S_i \) (sample 629-1). The zoning can be distinguished on high contrast back-scattered electron (BSE) maps (Figure 6c).

5.1.3 | D3 and D4 deformation phases

The S2 fabrics were then involved in S/C/C' structures during the D3 deformation phase (Figure 4). In sample S16-2, these structures rotate the porphyroblasts with respect to S2 in ‘dextral’ non-coaxial flow and wrap the S2 fabric around the porphyroblasts (Figures 4a and 5c). As a result, \( S_i = S_2 \) in the rim of garnet) appears oblique to the external foliation in the matrix (\( S_n \)).

D3 resulted in the main mylonitic foliation S3 separating the polymineralic S2 layers (Figures 4 and 5) and closed to isoclinal folds (F3) with subhorizontal fold axial surfaces (Figures 2 and 3). The S3 fabric, which is parallel to the axial surface of F3 and corresponds to C in the S/C/C' structure, is a subhorizontal to gently dipping mylonitic fabric. This fabric defines a spaced foliation (\( \mu \mathrm{m} \) thick) that is pervasive in the mica schist and can be observed from thin section to outcrop scale (Figures 3–5; Bender et al., 2018). The mineralogical assemblage that defines S3 is dominated by biotite but also includes white mica, plagioclase and epidote group minerals (commonly zoned allanite–epidote), as well as accessory phases such as opaque minerals including ilmenite and pyrite, and...
FIGURE 5  Legend on next page.
zircon (Figure 5). The mylonitic fabric S3 (shear band C) is also defined by recrystallized or newly formed minerals such as quartz (Figure 5i,j). They form various ribbon-like structures, either as quartz newly developed around plagioclase and garnet porphyroblasts (mantle porphyroblasts, sample 585-5) or constitute independent regularly spaced (μm to mm thick) bands of recrystallized quartz grains (Figure 5i,j). The latter fabric exhibits amoeboid grains, partly consisting of several subgrains, which suggests combined grain boundary migration and subgrain rotation recrystallization, respectively. Rare bulging recrystallization can also be observed (Figure 5i,j). (k, l) Locally, the mylonitic foliation S3 is affected by thin discrete C'-type shear bands defined mainly by chlorite and white mica. These C'-type shear bands may have formed originally under epidote-amphibolite facies conditions, like the S3 mylonitic fabric corresponding to C-type shear bands, and subsequently been reworked at lower grade conditions.

deformation phase. The S2 fabric characterized by amphibolite-grade deformation is weakly to strongly overprinted by D4. Biotite and white mica layers are to different extent chloritized (samples 582-2, 585-5, 595-6, 615-2, 641-6 and 695-5; Figures 4c and 5e). Fractures in garnet are filled by white mica and chlorite (Figure 5c). Some of the porphyroblasts are overgrown by chlorite (sample 582-2).

5.1.4 | Petrography

Mineral inclusions in garnet are represented by quartz, ilmenite, plagioclase, phengite, biotite, chlorite, apatite, rutile and allanite–clinozoisite. Ilmenite inclusions are found throughout the garnet porphyroblasts, whereas rutile is located mainly in the late D2 garnet rims. Additionally, the rutile–ilmenite assemblage is stable within the same garnet zone.

Polymineralic layers of mainly white mica II and biotite II together with quartz and plagioclase II wrap the garnet porphyroblasts and define the external S2 foliation (Figure 5c). Pressure shadows contain all major matrix minerals, plagioclase, white mica, biotite and quartz. Some fractures in garnets are filled by elongated chlorite, which also marks S3 fabric and C' bands (Figure 5c–k).

Pyrite forms elongated porphyroblasts, which range in size from 0.2 × 0.6 to 0.5 × 2.2 mm (Figure 5a,b). In addition, small (<10 μm) pyrite grains are also found in the matrix. Inclusions of rutile, ilmenite, plagioclase (Na–Ca), zoned white mica, biotite, quartz and monazite surrounded by a corona with allanite–zoisite, apatite and zircon, as well as polyphase inclusions of white mica–biotite–rutile, are present in pyrite porphyroblasts (Figure 5b).
FIGURE 6 (a–f) Back-scattered electron (BSE) image and X-ray maps of (a–c, e–g) Ca, Mg and Mn in garnet porphyroblasts and (g) Na, Al and Ca in plagioclase porphyroblasts for representative samples. Core–rim boundaries are marked. Values given in blue represent Ti concentration in quartz inclusions, whereas red values show the inclusion pressure of quartz (in GPa). Sample and garnet IDs are given in the upper right corner of the BSE images.
Rutile is found in the matrix and as inclusions in plagioclase, garnet and pyrite. Rutile was rimmed by ilmenite, indicating a decompression took place during exhumation.

Ilmenite is found in the matrix, as inclusions in garnet, plagioclase and pyrite, and is intergrown with white mica.

Zircon occurs in the matrix and as inclusions in white mica, biotite and garnet. One of the matrix zircons contains inclusions of apatite.

Calcite (<10 μm) is present as an accessory mineral in the rock matrix and as inclusions in plagioclase.

Matrix-hosted apatite is commonly elongated and ranges in size from 20 μm to 2 mm (length) (Figure 5c). Someapatite grains also contain inclusions represented by phengite, biotite, plagioclase and quartz. Apatite is found as inclusions in garnet, pyrite and zircon.

The epidote group minerals are represented by allanite, epidote and clinozoisite and are found in the rock matrix and as inclusions in garnet, plagioclase and pyrite. They show pronounced zoning; the cores are the richest in rare earth elements (REE) and the rims are depleted in REE. The relic monazite is partially replaced by apatite, allanite and clinozoisite (Figure S2a). The other microtextures observed are apatite–allanite–clinozoisite, allanite–clinozoisite and apatite-thorite–epidote (Figure S2b). Quartz, white mica and biotite are the inclusions found in allanite.

5.2 Mineral chemistry and chemical maps

The mineral chemistry is given only for sample 516-2 (Tables 2 and 3) as this sample was used for the GBMP geobarometric calculations and phase equilibrium modeling. As mentioned above, this sample preserves well-equilibrated mineral assemblages and lacks well-developed features of late retrogression.

5.2.1 Mineral chemistry

Sample 516-2 is composed of plagioclase (~60–63 vol.%), biotite (~20–22 vol.%), white mica (~5–7 vol.%), pyrite (~3–5 vol.%), allanite–epidote–clinozoisite (~3–4 vol.%), quartz (~2–3 vol.%), garnet (~1 vol.%) and a minor amount of K-feldspar, rutile, ilmenite, zircon, monazite, apatite and calcite. Different types of plagioclase are present. Plagioclase I corresponds to syn-S1 to intertectonic porphyroblasts (Figure 5a) ranging in size from 1 to 8 mm within which inclusions of biotite, white mica, quartz, rutile, ilmenite, apatite, pyrite and allanite–clinozoisite are found. This plagioclase is heterogeneous; in the BSE images, lighter porphyroblasts (Ab85–87An12–14Or0.37–0.44) contain darker patches that represent more Na-rich plagioclase (Ab92–93An7–8Or0.26–0.33) and quartz (Figure 5a and Table 2). Within the S2 foliation, we recognize plagioclase II, which is Na poorer than plagioclase I (Ab81–85An15–18Or0.32–0.71), together with biotite II, white mica II, quartz, ilmenite and apatite (Figure 5a,b). Plagioclase with the latter composition is also found as inclusions in white mica.

Most of the white mica defines the S2 foliation (white mica II; Figure 5). It is homogeneous and represented by muscovite (Si = 3.10–3.21 apfu, X_Mg = 0.35–0.46). However, some phengite grains (white mica I) found as inclusions in plagioclase porphyroblasts I are zoned with Si-rich cores and Si-poor rims (BSE-dark cores and BSE-light rims, respectively). The Si content in the BSE-dark cores ranges from 3.31 to 3.38 apfu and X_Mg = 0.56–0.62 (Table 3). The BSE-light rims have the same composition as the matrix muscovite. One white mica inclusion has also been found in apatite (Si = 3.23 apfu, X_Mg = 0.52). Some of the white mica also define the S3 mylonitic foliation (white mica III [Si = 3.13–3.16 apfu, X_Mg = 0.40–0.43]) and probably the C’-type shear band (Figure 5k,l).

Structurally, three different types of biotite are represented. Biotite I is found as inclusions in plagioclase I, biotite II is associated with S2 and biotite III is associated with S3 (Figure 5). Despite its microstructural position, biotite is chemically homogeneous, with X_Mg = 0.52–0.54 and Ti = 0.10–0.12 apfu (Table 3).

Syn-D2 garnet porphyroblasts range in size from 0.8 to 3 mm and are rich in mineral inclusions. Almandine and grossular slightly increase from core to the rim whereas spessartine and pyrope show opposite trends. The composition varies from Alm44–47Sp81–90Gr85–15Pr86–7 for garnet core through Alm46–48Sp23–45Gr87–13Pr86 for garnet mantle to Alm27–49Sp25–26Gr80–83Pr5–6 for garnet outer rim (Table 2 and Figure 6b).

There is only a single grain of K-feldspar (Or99.37Ab0.61An0.02; Table 2) found in this sample. Additionally, K-feldspar forms exsolutions in plagioclase, parts of which are rich in BaO.

5.2.2 X-ray chemical maps

X-ray chemical maps of Ca, Mn and Mg reveal complex zonation in garnet porphyroblasts, which allow determining two main types of garnet and support the microstructural study (Figure 6). The first group, whose growth occurs syn-D2 to late D2, exhibits double rims (referred to as ‘inner’ and ‘outermost’ rims) surrounding the cores (Figure 6a–e). The core–rim boundaries are commonly
irregularly shaped and embay the chemical zoning of the garnet cores. The garnet cores are distinguished by high Mn concentrations with typical prograde zoning profiles that display a progressive decrease towards the mantle and also through a sharper boundary towards the Mn-poorer outermost rim (Figure 6a–e). Magnesium concentration, which is relatively low and homogenous, does not display a clear trend. Garnet in the samples 516-2 and 615-2 show a slight decrease of Mg from core to rim (Figure 6a–c), whereas garnet in the deeper located samples (683-4 and 690-6) show the opposite trend (Figure 6d,e). Calcium zoning shows substantial variability between the different samples but commonly reciprocates Mn with a general increase towards the outermost rim and smaller scale fluctuations that mark the core–mantle–outermost rim boundaries (Figure 6a–c). Strongly heterogeneous and patchy calcium zoning is observed in some samples (683-4 and 690-6; Figure 6d,e). The garnet outermost rims are distinguished by high Ca concentrations and Mn depletion.

The second group of garnet, observed in the sample 629-1 (Figure 5f), is characterized by post-D2 smaller euhedral porphyroblasts, which display itself two generations of garnet with a sharp change in major element chemistry that follows the garnet I–garnet II boundary (Figure 6f). Garnet I is rich in Ca and depleted in Mg and shows relatively high levels of Mn. The Ca pattern shows the dissolution of garnet I before the growth of garnet II. Garnet II displays a well-developed bell-shaped zoning with low Ca and high Mn at the boundary with garnet I (Figure 6f). Ca increases towards the garnet II rims, and Mn shows the opposite trend. Magnesium in garnet II is homogeneous and at lower levels in comparison with garnet I.

X-ray chemical maps of Na, Al and Ca collected in one plagioclase porphyroblast (sample 615-2) show an albite-dominated inclusion-rich core and an inclusion-free rim enriched in Ca (Figure 6g).

5.3 | Thermobarometry

Quartz inclusions occur in cores, mantles and outermost rims of garnet hosts that show compositional zoning. Hence, titanium concentration and Raman spectroscopy measurements in quartz inclusions were sorted in three populations based on garnet zoning: (1) quartz inclusions in garnet core, (2) quartz inclusions in garnet mantle and (3) quartz inclusions in garnet outer rim (Figures 6 and 7).

5.3.1 | Titanium concentration in quartz inclusions

The titanium concentrations \( n = 55 \) in quartz inclusions \( n = 32 \) from syn-D2 garnet \( n = 8 \) in samples 516-2, 615-2, 641-6 and 690-6 range from \(~5\) to 21 ppm (Table S2 and Figure 8a). Several outliers are caused by secondary fluorescence of nearby phases or partial re-equilibration (recrystallization of quartz grains in contact with matrix or in fractures). Titanium concentrations were not measured in quartz inclusions from post-D2 garnet, as observed in sample 629-1. In the host syn-D2
garnet, the Ti concentration varies from 134 to 2063 ppm. Larger quartz inclusions were measured along transects that radiate from the inclusion into the surrounding garnet. In most cases, individual quartz inclusions do not show any noticeable spatial variations in Ti concentrations (max \( \leq 3 \) ppm). However, for those showing a significant spatial variation in Ti concentration, that is, \( >4 \) ppm, the lowest values were considered as outliers and excluded from the average calculation. Titanium concentrations in quartz inclusions are correlated neither with inclusion size nor with the Ti concentration of adjacent host garnet. The quartz inclusions in the syn-D2 garnet core \( (n = 15) \) yield a Ti average of \( 11.5 \pm 4 \) ppm (Figure 8a). Those measured in the mantle of syn-D2 garnet \( (n = 12) \) yield a lower average of \( \sim 8.8 \pm 3 \) ppm, whereas Ti concentration of quartz inclusion in the syn-D2 garnet outermost rim \( (n = 12) \) shows a slightly higher average of \( 14.4 \pm 5 \) ppm (Figure 8a). TiQ temperatures were calculated based on the average of Ti concentrations of each garnet zone (core, inner and outermost rim) between 0.5 and 1.5 GPa using the solubility equation of Thomas et al. (2010) (Figure 9). In order to determine where TiQ isopleths lie in the P–T space, the activity of \( \text{TiO}_2 \) must be known or assumed. The textural position of rutile and ilmenite was used to determine the TiO\(_2\) activity in this study. For the peak pressure conditions, the \( a\text{TiO}_2 \) equals 1 and for the lower P–T conditions (prograde P–T path), a range of \( a\text{TiO}_2 \) was taken \( (a\text{TiO}_2 = 0.8, 0.9, 1.0) \). The latter values were supported by the Ti activity models for pelitic rocks calculated by Ashley and Law (2015) (see Figure 9).

5.3.2 | Elastic geothermobarometry

Raman spectroscopy was performed on 35 quartz inclusions from syn-D2 garnet \( (n = 16) \) in nine different samples (516-2, 582-2, 585-5, 595-6, 615-2, 641-6, 683-4, 690-6 and 691-2) and from post-D2 garnet \( (n = 2) \) in sample 629-1 (Table S3). Figure 8b shows the \( P_{\text{inc}} \) (inclusion pressure) of inclusions from garnet cores, mantles and rims. Quartz inclusion \( (n = 5) \) from the syn-D2 garnet core yields an inclusion pressure average of \( \sim 0.13 \pm 0.02 \) GPa (Figure 8b). Those \( (n = 12) \) measured in the mantle of syn-D2 garnet yield a higher average of \( \sim 0.19 \pm 0.02 \) GPa. Finally, the inclusion pressure average calculated from quartz inclusion \( (n = 16) \) located in the outermost rim of syn-D2 garnet shows an even higher average of \( \sim 0.30 \pm 0.02 \) GPa (Figure 8b), suggesting a progressive increase of the entrapment pressure during the garnet growth process and thus during the D2 deformation phase. In the sample 629-1 (Figure 6f), quartz inclusions \( (n = 2) \) from post-D2 garnets show the opposite trend and the range of inclusion pressure is much lower. Indeed, a quartz inclusion that is located in the post-D2 garnet core of garnet I yields an inclusion...
pressure of 0.09 ± 0.12 GPa whereas the one located in the rim corresponding to garnet II yields an inclusion pressure of 0.07 ± 0.08 GPa (Figure 6f). Entrainment isomekes were calculated for the subset of 35 quartz inclusions between 400°C and 800°C using the software EoSFit-Pinc (Angel et al., 2017) and EntraPT (Mazzucchelli et al., 2021) (Figure 9).

5.3.3 | QuiG barometry and TiQ thermobarometry

The elastic model for quartz inclusions in garnet was combined with a TiQ solubility model (Thomas et al., 2010) and applied as a QuiG and TiQ thermobarometer (QuiG–TiQ) to estimate P–T conditions of garnet crystallization and quartz inclusion entrainment (Gonzalez et al., 2019; Johnson et al., 2021). The intersection of average isomekes with average TiQ isopleths respectively associated with the three distinct chemical zones of syn-D2 garnet reveals P–T estimates of a three-stage garnet growth occurring along a prograde path (Figure 9). Metamorphic conditions (Ms1) recorded by quartz inclusions in garnet cores range from 0.6 to 0.73 GPa and from 460°C to 500°C. Those recorded by the quartz inclusions in garnet mantles (Ms2) range from 0.75 to 0.85 GPa and from 490°C to 520°C. Quartz inclusions record peak metamorphic conditions (Ms3) in garnet outermost rims and range from 1.13 to 1.18 GPa and from 590°C to 610°C. Ms1 and Ms2 are relatively close in the P–T space and could be considered as one single continuous event, whereas Ms3 constitutes another stage of garnet growth. Temperature estimates have an uncertainty of ±24°C, derived from the calibration method and the 2σ of the measured Ti content. Pressure estimates have an uncertainty of ±0.3 GPa for a single analysis, due to the uncertainty in measuring the Raman peak positions (see Gilio, Angel, & Alvaro, 2021; Gilio, Scambelluri, et al., 2021, for a full discussion of error propagation in elastic geobarometry).

5.3.4 | Conventional thermobarometry (GBMP barometry)

The assemblage outermost rim of garnet + biotite II + muscovite II + plagioclase II contemporaneous with the peak P–T conditions is perfectly suitable for the GBMP geobarometer (Wu, 2015). Errors related to the input temperature and analytical measurements of mineral chemistry provide the total error of ±0.12 GPa. Calculations that use a suitable mineral assemblage (assuring the
equilibrium between the chosen minerals) for multiple minerals pairs (cf. Tables 2 and 3) and considering reduced conditions (Fe$^{3+}$(Fe$^{2+}$ + Fe$^{3+}$)) (Bt) = 0.116; Fe$^{3+}$(Fe$^{2+}$ + Fe$^{3+}$) (Grt) = 0.03) yielded consistent results with an average of 581°C and 0.93 GPa (Figure 10 and Table S5).

5.3.5 Thermodynamic modelling

Phase equilibrium modelling was performed on the best preserved sample 516-2 to obtain the peak P–T conditions. In this sample, the microstructural context of garnet porphyroblasts and the different mineral parageneses are well constrained. The bulk composition, together with a list of mineral assemblage and their modal amount are presented in Table 4. Compositional isopleths representing a measured range of composition in garnet, biotite, white mica and plagioclase were calculated and are shown in Figure 10.

The white mica–biotite–plagioclase–garnet–epidote–quartz–rutile–ilmenite stability field (Figure 10a) represents the observed paragenesis in the selected mica schist and lies between ~560–700°C and 0.75–1.15 GPa. The calculated compositional isopleths of spessartine and pyrope (for garnet core-to-rim composition), white mica II (Si apfu), biotite (Ti apfu) and plagioclase (entire Ab range) plot within the desired stability field (except the lowest values of spessartine and Si apfu in white mica).

The isopleths representing pyrope, the highest Si apfu in white mica II (Si = 3.19–3.21 apfu; matrix white mica defining S2 foliation) and Ti apfu in biotite II constrain the P–T conditions to ~0.78–0.98 GPa and 570–610°C. The spessartine points to higher P–T (~640–700°C at 0.96–1.15 GPa). However, the garnet constitutes less than 1 vol.% of the rock, and its compositional isopleths do not

Figure 10 (a, b) Pressure–temperature (P–T) pseudosection of the mica schist 516-2 from the Collisional Orogeny in the Scandinavian Caledonides (COSC-1) drill core calculated in the MnNCKFMASHTO system. (a) The stability field with the observed mineral assemblage is highlighted. (b) Isopleths represent garnet (Grt Sps, spessartine; Grt Prp, pyrope), plagioclase (Pl Ab, albite), white mica (Wm Si apfu, Si apfu in white mica) and biotite (Bt Ti apfu, Ti apfu in biotite) composition. The purple rectangle encompasses pyrope, biotite and white mica II with the highest Si apfu within the stability field representing the mineral assemblage of the rock. Diamond shows the results of the conventional geothermobarometry (garnet–biotite–muscovite–plagioclase [GBMP]). For details, see the main text.

Abbreviations with the lower case are used for pure phases and upper case for solid solutions: (1) Wm Wm Pl Cpx Am Grt sph qz; (2) Bt Wm Cpx Pl Am Grt sph qz; (3) Pl Am Ep Wm Bt Cpx Grt sph qz rt; (4) Cpx Pl Ep Bt Wm Pl Grt qz rt; (5) Wm Pl Ilm Bt Am Grt sph qz rt; (6) Pl Ilm Bt Wm Am Grt sph qz rt; (7) Pl Ilm Bt Am Wm Grt sph qz; (8) Wm Pl Cpx Am Grt sph; (9) Bt Wm Am Pl Grt Cpx sph; (10) Ep Cpx Pl Grt Wm qz; (11) Ep Pl Bt Wm Grt qz rt; (12) Pl Ilm Ep Bt Wm Am Grt sph qz; (13) Grt Bt Pl Ilm Am Ep Wm qz; (14) Pl Bt Ilm Wm Am Grt qz; (15) Ilm Grt Wm Bt Ep Pl qz; (16) Bt Ilm Ep Pl Grt Wm qz; (17) Bt Pl Ep Bt Grt Wm qz; (18) Pl Ilm Pl Bt Grt Wm qz; (19) Pl Pl Ilm Bt Grt sph qz; (20) Pl Ilm Wm Bt Grt Wm sph qz rt; (21) Pl Pl Pl Bt Grt Wm sph qz; (22) Pl Ep Bt Wm Pl Grt qz; (23) Wm Pl Ep Bt Grt Wm sph qz; (24) Chl Grt Wm Ep Bt Pl Cpx sph qz; (25) Pl Grt Bt Wm Ep Chl Ilm sph qz; (26) Grt Wm Ep Chl Ep Pl sph qz; (27) Grt Bt Ep Wm sph qz; (28) Pl Ep Bt Wm sph qz; (29) Wm Pl Wm Bt Ilm Grt sph qz; (30) Pl Pl Pl Bt Ilm Wm sph qz; (31) Pl Pl Pl Bt Ilm qz.
crosscut within the stability field of interest; thus, the P–T estimates based on garnet composition should be interpreted with caution. The GBMP results plot within the represented stability field (Figure 10b). Isopleths for white mica with the highest Si content (Si = 3.31–3.38; white mica found as inclusions in plagioclase) plot outside the stability field of the observed paragenesis. Instead, it plots in the higher P and lower T fields where amphibole and clinopyroxene (jadeite dominated) are stable (Figure S4).

Modal isopleths within the stability field that matches the observed mineral assemblage predict 2–12 vol.% of white mica, 18–20 vol.% of biotite, 58–68 vol.% of plagioclase, 0.2–0.8 vol.% of garnet, 1–6 vol.% of epidote, ~3 vol.% of quartz, 0.0–0.7 vol.% of ilmenite and 0.1–0.3 vol.% of rutile.

5.4 In situ and single-grain fusion 40Ar/39Ar geochronology

Seven analyses of white mica III from sample 595-6 (Figure 4c,d) were dated by single-grain fusion (n: 5) and in situ intragrain laser ablation (n: 2) techniques (Figure 11 and Tables 5 and S4). Six analyses provided >97.5% of radiogenic 40Ar and one analysis yielded 90.7%. The collective dates show a continuous distribution of 40Ar/39Ar dates from 429.4 ± 3.3 to 419.9 ± 9.1 Ma (1σ uncertainty) with two younger dates of 411.6 ± 7.3 and 408.1 ± 10.0 Ma (1σ uncertainty). No systematic patterns between the 40Ar/39Ar dates and the dating technique used are observed. A weighted mean of all seven dates provides 423.0 ± 5.0 Ma (2σ uncertainty; mean square of weighted deviates [MSWD]: 1.4).

Twelve analyses of white mica II and III from the sample 695-5 (Figure 5h) were also dated by single-grain fusion (n: 6) and in situ intragrain laser ablation (n: 6) techniques. Additionally, three in situ analyses of biotite III were conducted (Figure 11 and Tables 5 and S4). All analyses yielded >94.3% of radiogenic 40Ar. The resulting 40Ar/39Ar dates range between 435.0 ± 7.2 and 416.0 ± 7.6 Ma (1σ uncertainty) with one older outlier of 461.0 ± 7.6 Ma (1σ uncertainty). No correlations between the 40Ar/39Ar dates with the analysed mineral nor technique used are apparent. Excluding the outlier, a weighted average of the 40Ar/39Ar dates provides 423.4 ± 2.4 Ma (2σ uncertainty).

6 DISCUSSION

6.1 Multistage garnet growth

Texturally and chemically zoned syn-D2 garnet from the metasedimentary rocks of the LSN show evidence of multiple syntectonic growth and resorption periods. In spite of preserved chemical zoning, the patterns do not display a monotonic trend. They exhibit double rims (mantle and outermost rim) with a sharp boundary, which we interpret to reflect garnet growth during three distinct metamorphic stages both syntectonic with early D2 cores and mantle as well as late D2 outermost rims (Figure 12).
several samples, the two generations of garnet are preserved as strongly zoned cores and homogenous outermost rims (Figure 6). Irregular core–rim boundaries enclosing the core zoning profiles together with local truncation of curved S1 inclusion trails in the garnet interiors are consistent with resorption following core growth (e.g., Kohn & Spear, 2000; Tracy et al., 1976). The garnet rims display major element chemistries incongruous with those of the garnet cores notably Ca, which is consistent with growth of the garnet cores and rims during discrete crystallization events separated by an intervening period of garnet resorption.

Quartz-inclusion geothermobarometry (TiQ and QuiG) applied to the studied rocks of the LSN also indicates three distinct stages of garnet growth, including prograde (garnet core Ms1–garnet mantle Ms2) and peak metamorphic condition (garnet outermost rim Ms3), at epidote–amphibolite facies (460–520°C, 0.6–0.85 GPa) and amphibolite facies conditions (590–610°C, 1.13–1.18 GPa), respectively (Figures 9–12). In this method, the intersection of a QuiG isomeke and a TiQ isopleth corresponds to the P–T conditions of garnet crystallization and quartz inclusion entrapment. In contrast to what is expected for garnet growth occurring during changing metamorphic conditions, quartz inclusions, from core to rim, do not yield significantly different Ti concentrations from Ms1 to Ms3 (Figures 8, 9 and 12); therefore, the prograde garnet growth is mainly pressure dependent. However, the solubility of TiQ being dependent on αTiO₂, which is typically assumed to be fixed when quartz is grown or recrystallized in the presence of rutile, can vary more substantially if we consider the TiO₂ content in

<table>
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<th>Date ± 1σ (Ma)</th>
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Abbreviation: MSWD, mean square of weighted deviates.

*The single date of c. 461 Ma was excluded from the weighted average calculation.
quartz being buffered by other Ti-bearing phases (ilmenite and titanite). The textural position of Ti buffering phase was used to estimate the TiO$_2$ activity in this study as well as modelling of prograde TiO$_2$ activity (Ashley & Law, 2015). For the peak pressure conditions, Ms3, that correspond to the growth of the garnet’s outermost rim, aTiO$_2$ was fixed at unity (Figure 9) because garnet outermost rims contain quartz, rutile and ilmenite and quartz was clearly produced within the rutile stability field. For the lower P–T conditions, corresponding to the Ms1–Ms2 evolutionary stages of metamorphism and the growth of the garnet core and mantle, respectively, a range of aTiO$_2$ from 0.8 to 1 was modelled (Figure 9). Some garnet core and mantle also contain rutile inclusions, which suggests that TiO$_2$ from 0.8 to 1 was modelled (Figure 9). A minor difference in aTiO$_2$ attributed to the garnet growth stages Ms1/Ms2 and Ms3 will tend to increase the temperature gap between these three main stages and could explain why Ti concentrations were quite homogeneous along the prograde path despite of changing metamorphic conditions.

Additionally, conventional thermobarometry and thermodynamic modelling were performed on the best preserved mica schist sample 516-2 to obtain the peak P–T conditions. Those traditional techniques based on major element exchange and net-transfer only allowed the P–T conditions of the garnet’s outermost rim growth to be constrained as the core and mantle of garnet were not considered to be in equilibrium with the surrounding matrix. Within this sample, the mineral association garnet outermost rim + biotite II + muscovite II + plagioclase II and their representative isopleths were used to estimate the P–T conditions of final garnet-matrix equilibrium. Similar results were obtained with both techniques yielding lower pressure and slightly lower temperature amphibolite facies conditions 570–610°C and 0.78–1.00 GPa than those obtained with QuiG–TiQ geothermobarometry (Figures 10 and 12). The observed offset of pressure is relatively large (0.15 GPa) and outside the error of both techniques whereas the observed range of temperature is within the error (Figure 12). These discrepancies may be related to different factors, such as uncertainty in the reactive bulk composition from which the garnet nucleated (Johnson et al., 2021) and retrograde replacement or re-equilibration during exhumation process (Kohn & Spear, 2000). Thermobarometry and calculation based on thermodynamic modelling of phase equilibria are always subject to major assumptions as they use bulk rock geochemistry and/or mineral chemical compositions to reveal the ranges of stable P–T conditions where possible mineral assemblage can crystallize at chemical equilibrium. Instead, the mechanical and chemical equilibrium necessary for
applications of QuiG–TiQ thermobarometry require co-crystallization of quartz, garnet and rutile, which reduces system variance to three minerals and two univariant curves that intersect at one point in P–T space (Gonzalez et al., 2019). Titanium concentration and especially inclusion pressure obtained from quartz inclusion in the sample 516-2 are perfectly concordant with those obtained from the eight other samples in spite of lithological varietics, which suggest that QuiG–TiQ results are well constrained and provide a consistent P–T estimate for peak metamorphic conditions. The chemical re-equilibration of garnet may have occurred during the development of the regional mylonitic foliation, making the task of deciphering peak metamorphic conditions with traditional thermobarometric methods impossible. Nevertheless, information provided by those techniques is definitely not meaningless, as they deliver complementary results and necessary data to constrain the P–T evolution along the retrograde path.

A later period of garnet growth is also indicated by the occurrence of small post-D2 euhedral garnet grains in the samples 629-1 and 582-2 (Figures 5f and 6f). Two generations of garnet display a sharp change in major element chemistry that follows the garnet I–garnet II boundary. Micro-Raman spectroscopy was performed on quartz inclusion entrapped in garnet I (core) and garnet II (rim) and yields an inclusion pressure of 0.09 ± 0.12 and 0.07 ± 0.08 GPa, respectively (Figure 9b). Unfortunately, no measurement of titanium concentrations was collected in those quartz inclusions and, thus, the P–T condition corresponding to this later period of garnet growth could not be calculated. However, this garnet nucleation is interpreted as a retrograde growth occurring most probably at lower amphibolite facies conditions than Ms3.

6.2 | Garnet re-equilibration and mylonitic overprint during exhumation

The D2 deformation phase, associated with the development of S2 foliation and interpreted to record prograde to peak metamorphism through multistage garnet growth (Ms1 to Ms3) of the LSN, is then overprinted by the D3 deformation phase (Figure 12). The latter is characterized by a retrograde near isothermal decompression at ~620°C during the development of the main mylonitic foliation S3 (shear band C) that separates the polymetamorphic S2 layers (Figures 4, 5 and 12). The S2 fabric is indeed involved in S/C/C’ structures and, in some samples, these structures rotate the porphyroblasts with respect to S2 in ‘dextral’ non-coaxial flow and wrap the S2 fabric around the porphyroblasts (Figures 4a, 5c and 12).

Giuntoli et al. (2018) have demonstrated that fracturing, coupled dissolution–precipitation and dissolution–precipitation creep were responsible for the development of the mylonitic fabric in the amphibolites from the middle and lower portions of the COSC-1 borehole (Figure 3) at conditions of ~600°C, 0.75–0.97 GPa. The latter P–T estimate (dashed green circle in Figure 12) matches, within error, with the P–T estimates we obtained using conventional thermobarometry and thermodynamic modelling methods at 570–610°C and 0.78–1.00 GPa in the garnet-bearing mica schist (Figure 12; Ms4 stage). Moreover, the mica schist sample 516-2 used for the P–T calculation shows strong mylonitic overprint (Figure 4a, b) with thin discrete C’-type shear bands. Additionally, monazite observed in this sample is partially replaced by apatite, allanite and clinozoisite (Figure S2), which provides clear evidence of a dissolution–reprecipitation process. Apatite might have reprecipitated from Ca-rich fluids during decompression stage. Therefore, we suggest that a chemical re-equilibration of the mica schist might also have occurred during the development of the regional mylonitic foliation, causing major changes in the chemical composition of several key minerals used for P–T estimates including garnet. Outermost rim compositions of garnet porphyroblasts are commonly reset and record the incipient stages of the development of the mylonitic foliation corresponding to the fourth metamorphic stage (Ms4) instead of peak metamorphic conditions (Figure 12). Several studies (e.g., Dempster et al., 2019; Dempster, La Piazza, et al., 2017; Dempster, Symon, & Chung, 2017; Hames & Menard, 1993; Martin et al., 2011; Wolfe et al., 2021) characterized and explored the mechanisms of garnet replacement by fluid-mediated dissolution–reprecipitation reactions. The introduction of fluids during metamorphism can induce reaction processes that may partially or completely re-place anhydrous minerals like garnet. A coupled dissolution–reprecipitation process is expected to create a sharp compositional gradient at the interface between the parent crystal and the re-equilibrated domains (Geisler et al., 2007; Putnis, 2002). In this study, the sharp compositional boundary separating the garnet mantle from the garnet outermost rim could correspond to the interface parent crystal–re-equilibrated domains (Figure 6). However, no cloudy zones characterized by numerous fluid inclusions, as described by Wolfe et al. (2021), were observed in the studied garnet. Hence, additional investigations are needed to further support garnet coupled dissolution–reprecipitation process induced by fluid.

The third deformation phase D3 is generally weakly to strongly overprinted by later lower grade, greenschist-grade deformation (D4). Indeed, C’-type
shear bands and the S3 mylonitic fabric that corresponds to C-type shear bands were reworked at lower grade conditions (Ms5) but still with the same kinematics, when the LSN was juxtaposed with the Offerdal Nappe. Bender et al. (2018) and Merz et al. (2019) described a similar evolution; these two groups of structures display the same kinematics in the Offerdal-Seve shear zone: a top-to-the-ESE mylonitic epidote-amphibolite facies fabric that is overprinted by a top-to-the-ESE greenschist facies fabric. Giuntoli et al. (2020) have also well constrained these two stages with (1) the main mylonitic foliation formed at epidote–amphibolite facies conditions during the retrograde path from 600°C, 1.0 GPa to 500°C, 0.5 GPa and (2) the later overprint as a discrete, brittle-to-ductile C and C′ shear bands, developed at metamorphic conditions of 300–400°C and ~0.3 GPa (see dashed red square in Figure 12). The brittle character of D4 allowed preferential fluid influx along C/C′ planes, which led to chlorite and white mica crystallization.

Biotite and white mica from the samples 595-6 and 695-5 yield weighted mean 40Ar/39Ar populations of 423.0 ± 5.0 and 423.4 ± 2.4 Ma, respectively (Figure 11). Both samples originate from the lower part of the drill core (~1880 and 2500 m depth, respectively; Figure 3) and are affected by strong shearing and mylonitization as well as greenschist facies overprinting (Ms5–D4). The analysed mica are syn-kinematic with shearing and show strong evidence of recrystallization, which is considered to be a main driver for 40Ar loss in mica and resetting of the 40Ar/39Ar dates in greenschist facies conditions (e.g., Di Vincenzo et al., 2004; Hueck et al., 2020; Ring et al., 2007; Villa, 1998). Therefore, the 40Ar/39Ar ages of the deformed mica from the LSN are interpreted to resolve the timing of deformation along the basal shear zone of the LSN bracketing the interval between ~1880 and 2500 m depth of the COSC-1 borehole.

The deformation ages obtained reproduce previous work from the lower part of the drill core that resolved the main mylonitic foliation development to 417 ± 9 Ma using U–Pb dating on syn-kinematic titanite grains in the amphibolites (Giuntoli et al., 2020). The shearing occurred under epidote–amphibolite facies conditions along the retrograde path from 600°C and 1 GPa to 500°C and 0.5 GPa. At Åreskutan, the MSN was deformed at c. 424 Ma, constrained by in situ Th–U–Pb dating of monazite from granulite facies rocks (Majka et al., 2012). Moreover, several other studies (Bender et al., 2019; Dallmeyer, 1990; Dallmeyer et al., 1985; Grimmer et al., 2015; Gromet et al., 1996; Hacker & Gans, 2005) have shown that the tectonic contact between the MSN and LSN was active between c. 435 and 426 Ma. Furthermore, exhumation of the Upper Seve Nappe southwest of Åreskutan and the COSC-1 drill site has been dated to c. 418–415 Ma (Ma et al., 2022). Therefore, our new 40Ar/39Ar data, together with the previous studies, demonstrate that exhumation of the entire SNC package was coeval at different crustal levels and likely coincides with thrusting of the LSN onto the Offerdal Nappe (Gee et al., 2020) (= transition D3 to D4; cf. Figure 12).

### 6.3 Geodynamic significance of the LSN evolution

A key element to understanding the tectonic evolution of the LSN in west-central Jämtland is the orthogneiss that is hosted in the LSN metasedimentary rocks, for example, as exposed at the Så quarry (Li et al., 2021). The structural relationships between the orthogneiss and the calc-silicate host rocks in the quarry indicate that the orthogneiss protolith intruded prior to development of the main regional mylonitic foliation, which is most likely the equivalent of S3 that is detailed from the borehole samples. Therefore, the orthogneiss protolith intruded the LSN prior to exhumation from peak metamorphic conditions of 590–610°C and 1.13–1.18 GPa (i.e., pre-D3). Recent geochronology of the orthogneiss (leucogranite in Li et al., 2021) indicates an intrusive age at c. 468–470 Ma, whereas dating of white mica associated with S4 from borehole samples provided a deformation age of c. 423 Ma. The latter agrees with the results from previous works resolving the emplacement of the LSN (i.e., Bender et al., 2019; Giuntoli et al., 2020). As a result, the intrusion age of the leucocratic orthogneiss protolith provides an important constraint for the timing of peak metamorphism for the LSN, whereas the final ages of emplacement provide the minimum age bracket. Provided the wealth of evidence that (U)HP conditions were realized in the SNC in Jämtland at c. 460 Ma (Brueckner & Van Roermund, 2007; Fassmer et al., 2017; Grimmer et al., 2015; Majka et al., 2012), it can be assumed that the LSN, in the Åre Synform, also reached peak conditions at a similar time prior to the onset of exhumation. However, the tectonic setting prior to the pre-D3 emplacement of the orthogneiss remains enigmatic, yet it is crucial for understanding the overall tectonic evolution of the LSN in relation to the rest of the SNC, specifically the MSN. Evidence for subduction and high-grade metamorphism of the MSN at c. 482 Ma was reported by Walczak et al. (2022), which indicates that subduction of the Baltica margin had begun well before the intrusion of the orthogneiss in the LSN at c. 470 Ma (Li et al., 2021). This chronology requires the LSN to be located in a tectonic setting close to a subduction zone.
that would permit the intrusion of the orthogneiss protolith. Taking this additional information into account, two different tectonic models (Figure 13) have been proposed to reconstruct the possible tectonic scenarios that satisfy the developing state of knowledge for the SNC in west-central Jämtland between the early/middle Ordovician and late Silurian.

**FIGURE 13** Schematic geological 3D block diagram (a, b) and cross-sections (c, d) illustrating two possible tectonic evolutions of the Seve Nappe Complex in west-central Jämtland in the Scandinavian Caledonides, between the early/middle Ordovician and late Silurian. Model 1: (a) Initially, two detached slivers of Baltica belong to the Middle Seve Nappe (MSN) and the Lower Seve Nappe (LSN) rocks; (b) subduction of the outermost margin including the MSN rocks while the more proximal LSN rocks approached the subduction zone and undergo a substantial extension before their collision with the upper plate. Extension of the lower plate causes decompression melting of the deeper crustal levels and provokes the emplacement of the leucocratic orthogneiss protolith into the LSN; (c) introduction of the LSN rocks into the subduction zone reaching peak conditions; (d) exhumation and MSN emplacement onto the LSN. Model 2: (a) Initially, the LSN rocks are located on a detached sliver of Baltica as a piece of the upper plate involved in an arc-forearc system; (b) intrusion of the orthogneiss protolith at c. 470 Ma into the LSN host rocks as a direct consequence of subduction-related igneous activity in the arc-forearc area; (c) during ongoing subduction, the LSN rocks are scraped off from the arc-forearc into the subduction channel, proceeding to the peak metamorphic conditions; (d) subsequent exhumation and nappe-stacking similar to Model 1. The light purple plate refers to the oceanic crust and Island arc (Upper Allochthon, Köli Nappe Complex); the light blue refers to accreted sediments in the forearc basin. The white plate with cross igneous symbols refers to the upper and lower crust (Lower Allochthon and Middle Allochthon), and finally, the sedimentary cover is in orange and comprises Middle Allochthon rocks (Seve Nappe Complex and lower part). For further explanation, see the text in Section 6.3.
Model 1 assumes two detached slivers of Baltica corresponding to the MSN and LSN rocks for the more proximal one. They both belong to the lower plate of the subduction system. Subduction of the outermost margin (i.e., continent–ocean transition zone), including the MSN rocks, started in the late Cambrian, while the more proximal LSN rocks approached the subduction zone. The LSN rocks undergo a substantial extension before their collision with the upper plate, owing to pull from the trenchward part of the subducting plate (Gün et al., 2021). In this scenario, extension of the lower plate could cause decompression melting of the deeper crustal levels, that is, in the crystalline basement beneath the sedimentary basin of the LSN (Li et al., 2021), and provoke the emplacement of the leucocratic orthogneiss protolith into the LSN. The focus of the extension on the continental lithosphere embedded within the oceanic lithosphere might be due to the rheology and thickness differences between the oceanic and continental crust as demonstrated by Gün et al. (2021). As subduction proceeded, the rocks of the LSN were introduced into the subduction zone and reached peak conditions before being exhumed and overthrust by the MSN. Nevertheless, this model presents several limitations and raises some questions such as the following: What is the extension rate affecting the lower plate that is required to generate melting of the deep crust and can it be reached in this specific setting?

Model 2 also involves late Cambrian subduction of the Baltica margin, but instead of placing the LSN rocks on the lower plate during the Caledonian orogeny, the LSN rocks are located on a detached sliver of Baltica as a piece of the upper plate involved in an arc-forearc system (see example of the Andaman sea in Singh et al., 2013). The location of the LSN in this model is the result of hyperextension of the Baltica margin during Iapetus Ocean opening, tearing pieces of the Baltica margin into the Iapetus realm. This model requires the presence of a narrow sea between the sliver and proper Baltica. Subsequently, the formation of a subduction zone along the Baltica margin occurred between the LSN rocks and the higher structural positions of the SNC, that is, the MSN rocks. The intrusion of the orthogneiss protolith at c. 470 Ma into the LSN host rocks would have been a direct consequence of subduction-related igneous activity in the arc-forearc area. The location in an arc-forearc setting also explains the presence of copper sulphide mineralization within the LSN in west-central Jämtland (Helfrich, 1967). During ongoing subduction, the process of subduction erosion (Huene et al., 2004; Straub et al., 2020) scraped off the LSN rocks from the arc-forearc into the subduction channel, proceeding to the peak metamorphic conditions. Subsequent exhumation and nappe-stacking is similar to Model 1. As for the first model, several limitations can be highlighted including the exact location of the LSN rocks into the arc-forearc system. These rocks need to collect melt from subduction-related igneous activity while being involved in the subduction erosion process 10 to 20 Ma later.

According to epsilon-Hf values obtained from the zircon in leucocratic orthogneiss by Li et al. (2021), melt could have sourced from the crystalline basement rocks, perhaps from the Sveconorwegian orogeny. This would suggest that Model 1 might be a likely scenario where melt was produced by decompression of the crystalline basement during transient extension, rather than flux melting of the mantle wedge of the subduction zone, which is inferred in Model 2. However, if pieces of the Sveconorwegian basement underlying the LSN were also detached from Baltica during Iapetus Ocean opening, then it is possible that heating of the crystalline basement as an indirect result of mantle wedge melting may have been responsible for the production of the orthogneiss protolith.

As stated earlier, both these two tectonic models have limitations. Timing of the intrusion outlines the constraints on the tectonic model and, consequently, it is not possible to invoke a normal oceanic or partial continental subduction, which may explain MSN/LSN thrusting and juxtaposition. Indeed, such a scenario does not suit the initial position of the LSN, especially at 470 Ma when it is intruded by the 468–470 Ma leucogranites, if we want to integrate into our model all the existing and published data.

7 | CONCLUSIONS

Applying QuiG and TiQ thermobarometry to subducted garnetiferous metasedimentary rocks provides new insights into the LSN prograde evolution. Quartz inclusions in garnet have preserved the P–T conditions of three garnet growth stages, which were not decipherable by traditional methods. They correspond to prograde to peak metamorphic stages Ms1 to Ms3. During Ms1 and Ms2 stages, the rocks were metamorphosed under epidote–amphibolite facies conditions (460–520°C and 0.6–0.85 GPa). The peak metamorphic conditions at the amphibolite facies stage (Ms3) were recorded by the garnet outermost rim and constrained to 590–610°C and 1.13–1.18 GPa. Conventional thermobarometry and thermodynamic modelling yield complementary results and deliver necessary data to constrain the retrograde evolution (Ms4) of these rocks. The mineral assemblage of the Ms4 stage equilibrated at ~570–610°C and ~0.78–1.00 GPa under amphibolite facies conditions. This retrograde stage, occurring during decompression,
is interpreted as a chemical re-equilibration of the meta-sedimentary rocks associated with the development of the regional mylonitic foliation and the introduction of Ca-rich fluids. This multi-method approach yields complementary results consistent with previous estimates of the conditions of metamorphism derived from the LSN. However, as demonstrated here, utilization of inherited phases in thermobarometric studies may lead to spurious interpretations and requires careful monitoring whenever working in poly-metamorphic terranes.

The reported $^{40}$Ar/$^{39}$Ar ages of syn-kinematic white mica and biotite from the strongly sheared and mylonitized mica schists under greenschist facies overprinting (Ms5) define the timing of the emplacement of the LSN onto the Offerdal Nappe. This thrusting occurred at c. 423 Ma.

Two potential tectonic scenarios were formulated in this study to explain the obtained P–T results for the LSN rocks and place them in the context of the large-scale tectonic evolution of the whole SNC. However, additional research is needed to validate these models and to elucidate the tectonic evolution of the SNC in the Scandinavian Caledonides.

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CONFLICTS OF INTEREST

The authors declare that there are no competing interests.

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**SUPPORTING INFORMATION**

Additional supporting information can be found online in the Supporting Information section at the end of this article.

**Figure S1:** $\varepsilon_1 + \varepsilon_2$ vs $\varepsilon_3$ graph showing the strains of the measured quartz inclusions in garnet. The figure also includes the hydrostatic stress line calculated from the quartz elastic tensor (blue dashed line; Wang et al. 2016), the isotropic strain line (green dashed line) and the lines of equal inclusion pressure of quartz (in GPa) calculated using two different methods (black and red lines). The lines in black ($P_{inc}$) were calculated by converting the strains into stresses using the adiabatic elastic tensor at 1 bar by Wang et al. (2015) and considering the inclusion pressure as the mean normal stress ($P_{inc} = \frac{\sigma_1 + \sigma_2 + \sigma_3}{3}$). The strain of the measured inclusions is generally not hydrostatic and appears to be dispersed parallel to the lines of equal $P_{inc}$. This has been interpreted as a possible sign of non-hydrostatic stresses, acting near the inclusion at entrapment (Gilio et al., 2021-JMG). Nevertheless, Gilio et al. (2021)-AmMin showed that there is a good agreement of $P_{inc}$ and $P_{trap}$ even for inclusions under strong differential strain.

**Figure S2:** a) Relic of monazite partially replaced by apatite, allanite and clinozoisite. b) Apatite – allanite – clinozoisite, allanite – clinozoisite, and apatite - thorite – epidote

**Figure S3:** T-O$_2$ pseudosection of the mica schist 516-2 from the COSC-1 drill core calculated in the MnNCKFMASHTO system at constant pressure (P = 1 GPa). The stability field with the observed mineral assemblage is garnet + white mica + biotite + quartz + plagioclase + epidote + rutile + ilmenite.

**Figure S4:** Pseudosection in the P–T range capturing the Si range of the phengite with the highest Si apfu (3.31–3.38 apfu; phengite inclusion in plagioclase).

**Table S1:** Sample names and their International Geo Sample Number

**Table S2:** Measured Ti concentrations in quartz inclusions and respective garnet hosts. Values in light grey have been excluded from average calculation

**Table S3:** Raman spectroscopy data for thirty-five quartz inclusions

**Table S4:** Compilation of in-situ and single-grain fusion $^{40}$Ar/$^{39}$Ar geochronology of white mica and biotite

**Table S5:** Results of garnet - biotite – muscovite - plagioclase thermobarometry


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