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Low-pressure isobaric cooling metamorphic *P–T* paths from Botnnuten in the southern Lützow-Holm Complex, East Antarctica

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Abstract

This paper reports the metamorphic texture of cordierite megacrysts and the metamorphic *P–T* path of a newly exposed section of gneiss in East Antarctica. We used mineral textures and pseudosection modeling to reconstruct the metamorphic *P–T* path of cordierite- and spinel–garnet-bearing gneisses from Botnnuten, an isolated nunatak located ~60 km from the southern edge of Lützow-Holm Bay in East Antarctica. The gneisses underwent low-*P* granulite-facies metamorphism at 5.0–6.1 kbar and 850 ± 20 °C followed by isobaric cooling. The isobaric cooling path implies long residence in the middle to shallow crustal level without rapid exhumation. This contrasts with the widely recognized clockwise *P–T* path of basement rocks of the Lützow-Holm Complex. The rocks at Botnnuten have long been considered part of the Lützow-Holm Complex based on their petrographical features and geothermobarometric data. However, the present results, combined with a reevaluation of available data, indicate the metamorphic history of the Botnnuten gneisses is more comparable to that of the Yamato Mountains, located southwest of the study area.

Keywords East Antarctica · Lützow-Holm Complex · Metamorphic *P–T* path · Isobaric cooling

Introduction

The evaluation of metamorphic *P–T* paths is critical for reconstructing tectono-thermal events and the tectonic setting of metamorphic rocks. During exhumation, the *P–T* path of high-grade metamorphic rocks typically shows isothermal decompression (ITD) or isobaric cooling (IBC). It has been hypothesized that the former is

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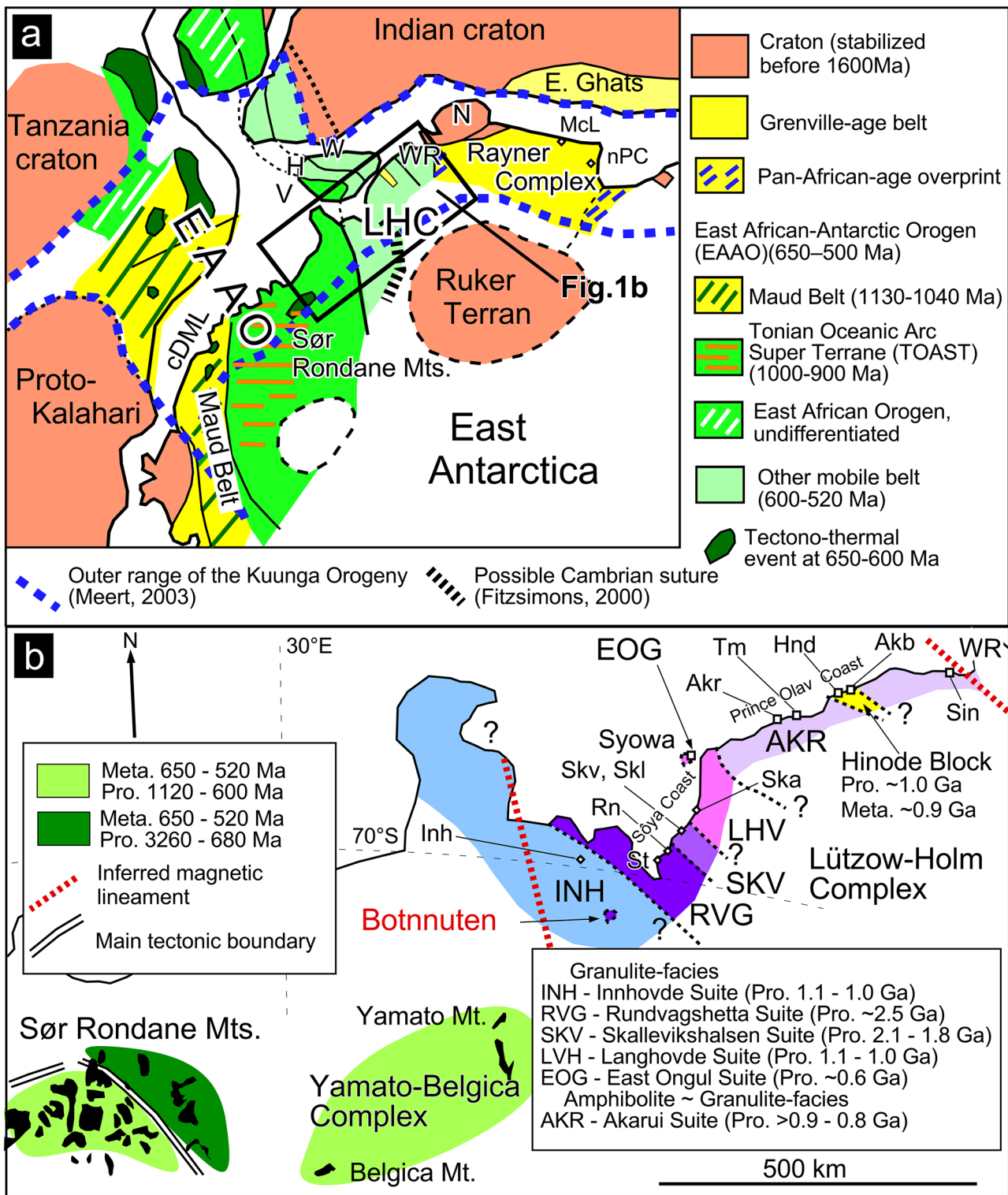6 The Hokkaido University Museum, Hokkaido 060-0810,
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due to the rapid erosion or uplift of metamorphic rocks in response to extensional tectonics after continental collision, and that the latter is due to magmatic loading in an extensional setting (e.g., Bohlen 1987, 1991; Harley 1989; Brown 1993). In many cases, IBC has been shown to occur during the retrograde stage of counter-clockwise *P–T* paths (Harley 1989; Bohlen 1991), but hybrid *P–T* paths showing early ITD and late IBC have also been reported (e.g., White et al. 2002; Baba et al. 2019). In general, many metamorphic terranes show evidence of ITD, which is considered to reflect the rapid exhumation of lower crustal rocks due to orogenic collapse or isostatic rebound of over-thickened crust formed by collisional orogeny (England and Thompson 1984; Thompson and England 1984; Harley 1989; Brown 1993).

The Lützow-Holm Complex (LHC) in East Antarctica is an Ediacaran–Cambrian orogenic belt that yields U–Pb zircon ages of 600–520 Ma (Shiraishi et al. 1994, 1997). The LHC was located in the center of the Gondwana supercontinent, which was formed by the global scale orogenesis involved convergent tectonics, which is termed as the East Africa–Antarctica Orogen (EAAO; Fig. 1a; Jacobs and Thomas 2004) at 650–500 Ma, or two separate East Africa Orogen (EAO) events (Stern 1994) at 650 ± 50 Ma and the

Fig. 1 (a, b)

subsequent Kuunga Orogeny at 550 ± 50 Ma (Meert 2003). The LHC records a clockwise metamorphic *P–T* path involving ITD (Yoshimura et al. 2008), as is evidenced by textures related to the breakdown of garnet to form orthopyroxene and plagioclase symplectites (e.g., Santoshi and Yoshida 1992; Takahashi and Tsunogae 2017; Takamura et al. 2020). The LHC is typically described as a collisional belt that was active during 600–500 Ma and extends through Sri Lanka,

South India, and Madagascar (e.g., Shiraishi et al. 1994, 2008; Meert 2003; Boger et al. 2015; Osanai et al. 2016; Durgalakshmi et al. 2021). Detailed metamorphic studies of the LHC have been conducted at several exposures along the Sôya Coast (Fig. 1b; e.g., Rundvågshetta: Kawasaki et al. 2011, Hiroi et al. 2019; Skallevikshallsen: Kawakami et al. 2016), revealing peak *P–T* conditions of 13–15 kbar and ~ 1040 °C, and that ITD cooling was involved with the interaction of a Cl-rich fluid (Kawakami et al. 2016) and the supercooling of melt inclusions (Hiroi et al. 2019). The LHC plays a crucial role in expanding our knowledge of a global geological structure beyond the orogenic belt. Moreover, it enhances our understanding of metamorphic processes and geofluids in the middle to lower crust.

Previous studies revealed that the metamorphic grade of the LHC shows a progressive increase from amphibolite to granulite-facies toward the southwest from Sinnan Rocks to Rundvågshetta (Fig. 1b; e.g., Hiroi et al. 1991; Yoshimura et al. 2008). Regional metamorphism of the entire LHC was considered to occur from Ediacaran to Cambrian (c. 600–520 Ma) (Shiraishi et al. 1994, 2008). These observations are inconsistent with recent reports of a local ultrahigh-temperature (UHT) thermal event of ~ 11–12 kbar and ~ 900–920 °C at Akarui Point (Fig. 1b)(Iwamura et al. 2013) and *P–T* conditions of ~ 11 kbar and ~ 830–850 °C estimated by Zr-in-rutile geothermometry and kyanite/sillimanite stability at Akarui Point, Skarvsnes, and Skallen (Suzuki and Kawakami 2019). Moreover, exposures of amphibolite-facies rocks at Akebono Rock in the eastern part of the LHC yield older (Tonian) metamorphic ages (Baba et al. 2021, 2022; Kitano et al. 2023) and kyanite-bearing pelitic gneisses preserve evidence for a counter-clockwise *P–T* path at Tenmondai Rock (Fig. 1b) (Baba et al. 2023), suggesting further investigation of the LHC is needed.

As outlined above, the metamorphic history of the entire LHC is complex and poorly understood due to isolated exposures surrounded by glaciers. During a survey of the 58th Japanese Antarctica Research Expedition, we visited the isolated inland nunatak of Botnnuten, located 60 km from the southern edge of Lützow-Holm Bay. We found coarse-grained cordierite-bearing gneisses containing large cordierite grains up to 18 mm in diameter, which had not been reported previously in the LHC. Although the *P–T* path of the LHC is characterized by ITD, the diagnostic ITD breakdown texture of garnet to form cordierite has been reported only in limited exposures that contain high-Mg lithologies (e.g., at UHT Rundvågshetta: Motoyoshi and Ishikawa 1997; Yoshimura et al. 2008; Hiroi et al. 2019), and cordierite itself is rare in the granulite-facies zone of the LHC. In this study, we report the petrography and model the metamorphic *P–T* paths of both cordierite- and spinel–garnet-bearing gneisses with the aim of reconstructing the regional tectono-thermal events in the LHC. The *P–T* paths show IBC trajectories that differ from those reported in previous works.

Geological setting

Lützow-Holm Complex

The LHC, an Ediacaran to Cambrian orogenic belt, occupies a vast area of about 500 km between the Western Rayner Complex (WRC) and the Yamato Mountains in East Antarctica (Fig. 1a and b; Shiraishi et al. 1994; 1997). Many studies have referred to the LHC as a 600–500 Ma continent collisional belt that also extends through Sri Lanka, South India, and Madagascar (e.g., Shiraishi et al. 1994, 2008; Meert 2003; Boger et al. 2015; Osanai et al. 2016).

The LHC consists of amphibolite- to granulite-facies metamorphic rocks and post-metamorphic intrusive rocks (Fig. 1b). The metamorphic rocks comprise quartzofeldspathic, intermediate to mafic, pelitic to psammitic gneisses, charnockites, granitoids, and calcareous and ultramafic rocks. The LHC is characterized by: 1) a progressive increase in metamorphic grade towards the southwest from Shinnan Rocks to Rundvågshetta (Fig. 1b; Hiroi et al. 1991); 2) a clockwise metamorphic *P–T* path inferred from kyanite relics within garnet porphyroblasts, which is considered to apply to the entire LHC (Hiroi et al. 1991; Yoshimura et al. 2008); 3) metamorphic ages of 600–520 Ma, obtained by sensitive high-resolution ion microprobe (SHRIMP) and laser ablation–inductively coupled plasma-mass spectrometry (LA-ICP-MS) zircon dating (Shiraishi et al. 1994, 1997; Dunkley et al. 2014, 2020; Takahashi et al. 2018; Takamura et al. 2018); and 4), which are inferred to data the timing of UHT metamorphism at Rundvågshetta (Fig. 1b), in the southwestern part of the region (Motoyoshi and Ishikawa

1997; Yoshimura et al. 2008; Kawasaki et al. 2011). As stated earlier, the latest data contradict the previous understanding of progressive metamorphism and timing in the LHC, indicating the need for revision.

Recently, the LHC has been subdivided into; 1) three tectonic units comprising a ca. 2.5 Ga continental crust, a ca. 1.0 Ga magmatic arc and a Neoproterozoic suture zone base on geochemical and geochronological data (e.g., Tsunogae et al. 2014; 2015; Takamura et al. 2018, 2020), or 2) six geological provinces known as the Innhovde Suite, Rundvågshetta Suite, Skallevikshalsen Suite, Langhovde Suite, East Ongul Suite, and Akarui Suite (Fig. 1b) based on protolith age and lithological characterization (Dunkley et al. 2020).
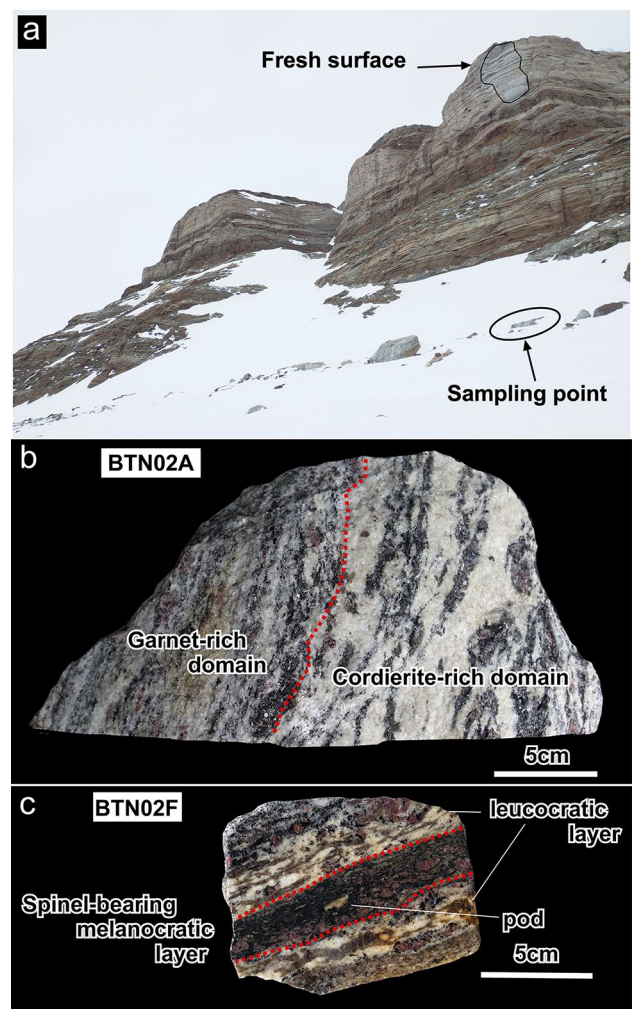
Botnnuten

Botnnuten (70°24'S, 38°01'E) is an isolated nunatak located ~170 km SSW of Syowa Station in East Antarctica. It comprises well-layered felsic and mafic gneisses, with subordinate calc-silicate rock, orthopyroxene-bearing orthogneiss (charnockite), and pelitic gneiss. The entire metamorphic sequence is ~400 m thick, and the foliation dips at ≤10° to the E or NE. The rocks in the middle-upper parts of the sequence are slightly more felsic than those in the lower part. Motoyoshi and Shiraishi (1985) reported metamorphic phase relations and geothermobarometry data that indicate the Botnnuten rocks reached P–T conditions of 5–6 kbar and 750–800 °C, and they concluded that the Botnnuten gneisses represent low-pressure rocks of the LHC. Dunkley et al. (2020) reported that charnockite and garnet-bearing semi-pelitic gneisses from Botnnuten yield protolith ages of 2472 ± 20 Ma and metamorphic ages of 590–550 Ma, respectively, and they interpreted the Botnnuten rocks as a klippe belonging to the Rundvågshetta Suite. During field investigations, we observed that a cliff on the northeastern side of the nunatak had collapsed, exposing a fresh surface (Fig. 2a). Samples of cordierite-bearing gneiss (SB17011802A, hereafter BTN02A) and spinel–garnet-bearing gneiss (SB17011802F, hereafter BTN02F) were collected from large, angular blocks (up to 10 m in length) located directly beneath the exposed cliff face. These samples are representative of the upper lithologies of the Botnnuten metamorphic sequence.

Analytical methods

Mineral composition

Prior to electron probe micro-analysis (EPMA), polished sections were covered with ~20 nm of electrically

Fig. 2 a Photograph of the northeast side of the Botnnuten. Higher side of the cliff has collapsed and shows fresh white surface; **b** and **c** Photographs of collected samples

conductive carbon to prevent charging. Pre-selected spots were analyzed in a JEOL JXA-8200 EPMA instrument. Analytical conditions were maintained at an accelerating voltage of 15 kV and a beam current of 12 nA for spot analyses and 150 nA for element mapping. The instrument was equipped with five wavelength-dispersive spectrometers (WDS), and all elements were measured at their respective Kα lines for 10 s on peak and 5 s on each upper and lower background. The used reference materials were wollastonite (Si), corundum (Al), periclase (Mg), hematite (Fe), spessartine (Mn), wollastonite (Ca), rutile (Ti), chromite (Cr), albite (Na), orthoclase (K), apatite (P, F). ZAF correction was applied to the analysis and corrected element contents were recalculated to weight percent. Total iron is reported as FeO_T.

Whole rock composition

To determine the whole-rock chemical compositions of the samples, rock slabs of each sample, about $50 \times 50 \times 10$ mm^3 for BTN02A and $25 \times 25 \times 10$ mm^3 for BTN02F, were ground in a stainless-steel mortar to a particle size of ~ 2 mm, and subsequently pulverized in a quartz ball mill using a Fritsch Pulverisette 6. Before fusion, all samples were heated at 1000 °C for 2 h to remove volatiles and oxidize ferrous iron. Powdered samples (1.8000 ± 0.0005 g) were thoroughly mixed with 2× excess Li-metaborate and fused to form a glass bead in an induction furnace using a NT-2000 Bead Sampler (Tokyo-Kagaku Co., Ltd.) by heating in several stages: first at 1200 °C for 2 min, and then swing heating at 1200 °C for 6 min. Whole-rock element contents were determined in a RIGAKU RIX 2000 X-ray fluorescence spectrometer instrument equipped with an Rh-tube operated at 50 kV and 50 mA. The loss on ignition (LOI) was determined gravimetrically by placing the sample powder in a furnace at 110 °C and 1000 °C, and weighing the sample at each temperature. The instrument operating conditions of the XRF spectrometers are documented by Kimura and Yamada (1996).

Phase equilibria modeling

To investigate the metamorphic conditions of a cordierite-bearing gneiss and spinel–garnet-bearing gneiss, a $Na_2O–CaO–K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O–TiO_2$ (NCKFMASHT) and NCKFMASHTO (NCKFMASHT–O_2) *P–T* isochemical phase diagram was computed for each sample based on free-energy minimization using Perple_X 6.8.7 software (Connolly 2005) and end-member thermodynamic data from Holland and Powell (2011) (filename: hp62ver. dat). The activity models of White et al. (2014) are used for orthopyroxene, biotite, garnet, cordierite, chlorite, mica, and staurolite, Wheller and Powell (2014) for sapphirine, White et al. (2002) for spinel, Fuhrman and Lindsley (1988) for feldspar, Holland et al. (2018) for melt and ideal solution models for ilmenite (IlGkPy). Details of these models and any modifications are given in the "Perple_X solution model glossary" and "Perple_X Updates" files, which are available at http://www.perplex.ethz.ch/perplex_updates.html.

Results

Petrography and mineral chemistry

Cordierite-bearing gneiss (BTN02A)

The cordierite-bearing gneiss (sample BTN02A) occurs as compositional layers within predominantly garnet–biotite–sillimanite gneisses (Fig. 2b). Sample BTN02A contains a cordierite-rich domain and a garnet-rich domain (Fig. 2b). The cordierite-rich domain (Fig. 2b) contains cordierite, biotite, sillimanite, K-feldspar, quartz, and plagioclase, with accessory zircon, ilmenite, and apatite. A small amount of garnet is also present. This domain is leucocratic with abundant feldspar-quartz layers that cross-cut the melanocratic lenses and layers. Cordierite megacrysts are up to 17 mm in diameter (Fig. 3a) and occur in localized clusters that are aligned with the gneissic banding (Fig. 2b). The cordierite is predominantly inclusion free, with rare biotite inclusions. The cordierite margins are partially replaced by biotite and subhedral–euhedral prismatic sillimanite (Fig. 3a). The X_{Mg} [molar Mg/(Mg+Fe)] value of cordierite decreases slightly from core (0.81) to rim (0.78) in a single analyzed crystal (Table 1 and Tables S1–S5). In places, garnet occurs alongside biotite at the margins of cordierite (Fig. 3b and c). Garnet crystals (up to 4 mm in size) are smaller than cordierite, and they have an irregular morphology with sharp crystal margins. The garnets have a narrow compositional range of pyrope (28–32 mol%), grossular (1.0–1.6 mol%), and spessartine (1.0–1.6 mol%; Fig. 4a and b). Biotite occurs predominantly at the cordierite margins and has X_{Mg} values of 0.55–0.62, and TiO_2 contents of up to 6.2 wt% (Fig. 4c). K-feldspar, quartz, and plagioclase occur in the matrix and are of variable size. K-feldspar is perthitic and shows lamellae textures such as stringlets, rods, and interlocking lamellae. Plagioclase and K-feldspar have near-uniform compositions of anorthite (An = 14–16 mol%) and orthoclase (Or = 89–93 mol%), respectively (Fig. 5a).

The garnet-rich domain (Fig. 6) contains garnet, cordierite, biotite, sillimanite, plagioclase, K-feldspar and quartz, with accessory zircon, ilmenite, and apatite. Garnet occurs as porphyroblasts up to 8 mm in diameter, which are partially replaced by biotite and rarely by cordierite (Fig. 6a). Garnet porphyroblasts show a compositional range of pyrope (28–40 mol%), grossular (1.4–1.7 mol%), and spessartine (0.6–1.1 mol%). The pyrope contents of garnet are higher in the garnet-rich domain than in cordierite-rich domain, and are highest in the garnet cores. Cordierite occurs as isolated euhedral grains in the matrix (Fig. 6b), secondary grains replacing garnet (Fig. 6a), inclusions within garnet (Fig. 6c), and as grains that are partially surrounded by garnet (Fig. 6d). Cordierite grains adjacent to the garnet margins have higher X_{Mg} values (0.79–0.80) than other cordierite grains (0.74–0.78). Biotite in the garnet-rich domain has higher X_{Mg} ratios (0.70–0.59) than biotite in cordierite-rich domains, and biotite inclusions in garnet have similarly high X_{Mg} ratios (up to 0.70) and moderate F contents (up to 0.52 wt%; Fig. 4c). Plagioclase shows little compositional variation (An = 14–15 mol%).

Table 1 Chemical compositions (EPMA results), and calculated mineral formulae and end-member fractions, for selected minerals

Sample		BTN02A(Crd-rich domain)					BTN02A(Grt-rich domain)			BTN02F		
Spot No.		47	7	2	18	15	99	123	67	32	159	28
		Grt	Crd	Bt	Pl	Kfs	Grt	Crd	Bt	Grt	Spl	Pl
Major oxides (wt%):	LLD	C marg		in Crd			rim	G marg	G marg	core	Spl in Sil	
SiO$_2$	0.02	38.92	50.00	36.30	65.12	64.82	38.25	49.71	36.19	37.75	0.06	61.38
TiO$_2$	0.03	0.01	0.00	6.13	0.04	0.07	0.03	0.03	5.62	0.00	0.00	0.00
Al$_2$O$_3$	0.02	21.89	33.28	16.38	22.41	18.99	21.62	33.68	15.74	21.56	58.43	24.50
Cr$_2$O$_3$	0.03	0.08	0.02	0.14	0.01	0.00	0.04	0.03	0.15	0.00	0.32	0.00
FeO^T	0.05	30.20	4.97	15.37	0.01	0.00	31.60	4.75	17.31	33.06	37.16	0.02
MnO	0.04	0.59	0.00	0.00	0.01	0.00	0.51	0.00	0.00	0.96	0.04	0.05
MgO	0.01	8.45	10.46	12.10	0.02	0.00	7.23	10.49	11.34	5.24	3.86	0.00
CaO	0.01	0.48	0.01	0.04	3.31	0.05	0.52	0.02	0.08	1.35	0.02	5.86
Na$_2$O	0.02	0.00	0.05	0.11	9.76	1.25	0.02	0.01	0.11	0.05	0.03	8.23
K$_2$O	0.01	0.01	0.02	9.66	0.16	15.07	0.01	0.01	9.69	0.01	0.00	0.20
ZnO	0.01	-	-	-	-	-	-	-	-	-	0.21	-
Total		100.63	98.81	96.22	100.85	35.42	99.82	98.73	96.22	99.97	100.12	38.86
Calculated mineral formulae (apfu)^a												
O		12	18	22	8	8	12	18	22	12	4	8
Si		3.003	5.037	5.392	2.846	2.976	2.999	5.007	5.432	2.991	0.002	2.720
Ti		0.001	0.000	0.684	0.001	0.002	0.002	0.002	0.634	0.000	0.000	0.000
Al		1.991	3.952	2.867	1.154	1.028	1.998	3.998	2.784	2.013	1.944	1.280
Cr		0.005	0.001	0.016	0.000	0.000	0.003	0.002	0.018	0.000	0.007	0.000
Fe^{3+}		-	-	-	0.000	0.000	-	-	-	0.013	0.047	0.001
Fe^{2+}		1.948	0.419	1.909	-	-	2.073	0.400	2.173	2.177	0.830	-
Mn		0.039	0.000	0.000	0.000	0.000	0.034	0.000	0.000	0.064	0.001	0.002
Mg		0.971	1.571	2.680	0.001	0.000	0.844	1.574	2.537	0.619	0.163	0.000
Ca		0.039	0.001	0.006	0.155	0.002	0.043	0.002	0.013	0.114	0.000	0.278
Na		0.000	0.009	0.030	0.827	0.112	0.003	0.003	0.032	0.008	0.002	0.707
K		0.001	0.003	1.830	0.009	0.882	0.001	0.001	1.854	0.001	0.000	0.011
Zn		-	-	-	-	-	-	-	-	-	0.004	-
Sum		7.998	10.992	15.413	4.994	5.003	8.000	10.989	15.476	8.000	3.000	4.999
X$_{Mg}$		0.33	0.79	0.58			0.29	0.80	0.54	0.22	0.16	
Calculated end-member fractions (mol%):												
Grossular		1.32					1.45			3.85		
Pyrope		32.40					28.20			20.81		
Spessartine		1.29					1.14			2.15		
Almandine		64.99					69.21			73.18		
X$_{An}$					0.16	0.00						0.28
X$_{Zn}$											0.00	

LLD Lower limit of detection, *FeO*^{T} represents total iron oxide, *Fe*$^{3+}$ is calculated based on charge balance

C marg in marginal part of cordierite (Crd), *G marg* in marginal part of garnet (Grt)

X_{Mg} = Mg/(Mg+Fe), Xan = Ca/(Na+Ca+K), X_{Zn} = Zn/(Zn+Mg+Fe^{2+})

^{a}Calculated based on the quoted number of O atoms per formula unit

Spinel–garnet-bearing gneiss (BTN02F)

Spinel–garnet-bearing gneiss (sample BTN02F) occurs as melanocratic layers within predominantly garnet–biotite gneiss and exhibits a grano-porphyroblastic texture. The melanocratic layer contains garnet, spinel, biotite, plagioclase, and K-feldspar, with subordinate corundum and sillimanite, and accessory zircon, ilmenite, rutile, and apatite. Quartz, plagioclase, and K-feldspar can be seen in the leucocratic layer adjacent to the melanocratic
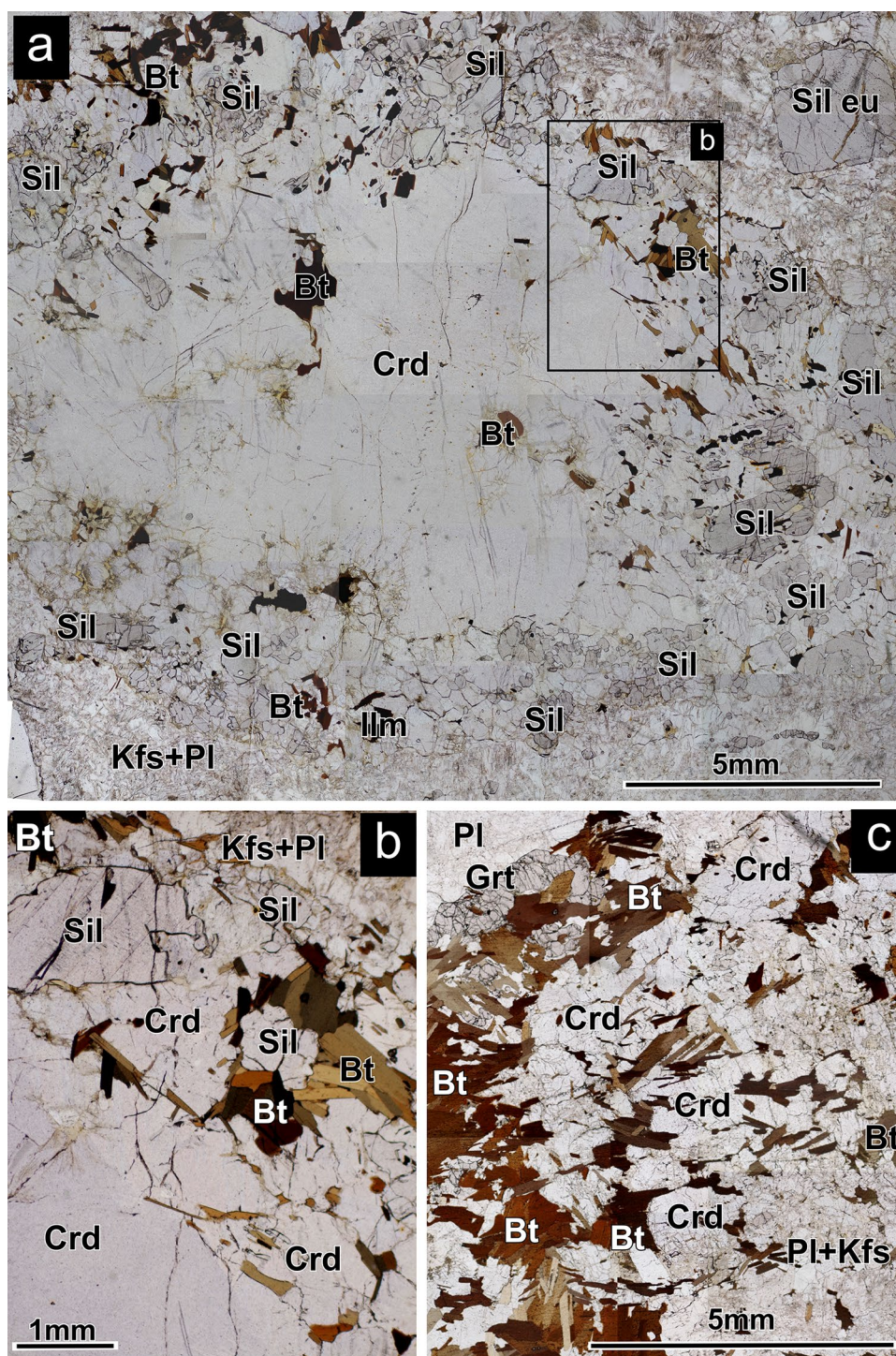
layer (Fig. 2c). The leucocratic layer partially invades the melanocratic layer and occurs as pods (Fig. 2c). In the melanocratic layer, garnet occurs as porphyroblasts up to 8 mm in diameter. The garnet contains scarce inclusions of biotite, spinel, ilmenite, and plagioclase that are small and rounded (Fig. 7a). Garnet locally occurs as overgrowths on spinel rims (Fig. 7a). Spinel occurs as granular crystals in the matrix of the gneiss (Fig. 7a

and b) and is weakly foliated, together with plagioclase. Small inclusions of spinel occur within garnet, sillimanite, and corundum (Fig. 7a, c, and d). Spinel grains in the matrix and included spinel have X_{Mg} values of 0.16–0.24 (Fig. 5b). Spinel ZnO contents are < 0.35 wt% and the Fe^{3+} content calculated by charge balance is < 0.07 apfu (O = 4). The biotite is finer-grained than the spinel and the garnet, and occurs as crystal aggregates (Fig. 7c) with
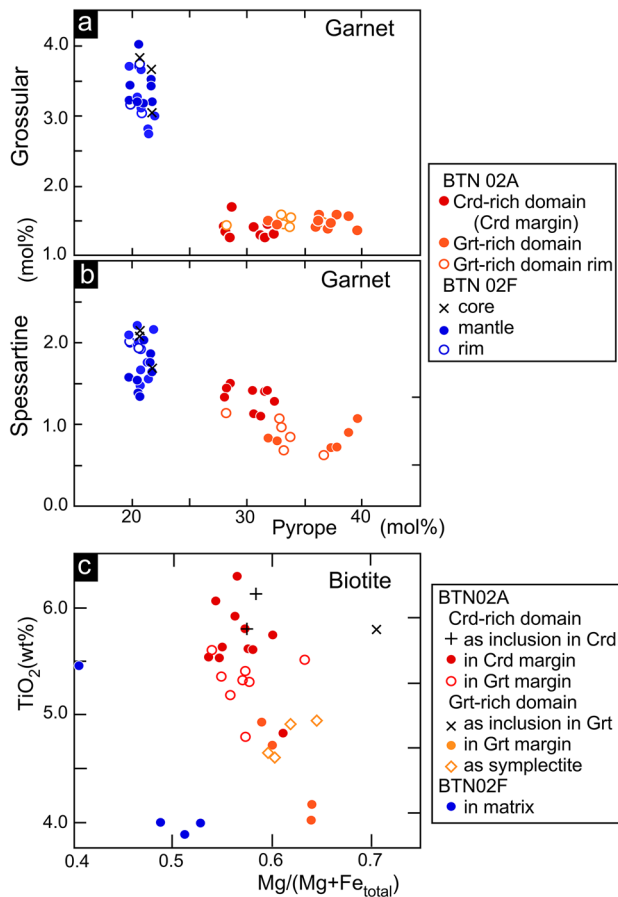
Fig. 4 **a** and **b** Compositional variations of garnet in terms of pyrope versus grossular and pyrope versus spessartine; **c** Compositional variations of biotite in terms of X_{Mg} versus TiO_2 contents. Grossular = Ca/(Mg + Fe + Ca + Mn) × 100, Spessartine = Mn/(Mg + Fe + Ca + Mn) × 100, Pyrope = Mg /(Mg + Fe + Ca + Mn) × 100

X_{Mg} ratios and TiO_2 contents of 0.40–0.53 and 3.9–5.5 wt%, respectively (Fig. 4c). Sillimanite and corundum occur together in the matrix with spinel. Corundum occurs as irregularly shaped crystals in the matrix and contains inclusions of spinel and ilmenite (Fig. 7b and c). Sillimanite occurs as euhedral to subhedral grains that contain inclusions of spinel and corundum (Fig. 7c). Locally, biotite replaces corundum and sillimanite at their grain margins. Sillimanite and corundum have Fe_2O_3 contents up to 0.43 wt% and 0.41 wt%, respectively. Plagioclase is uniform in size, has a granoblastic texture, and has a narrow range of anorthite contents (27–29 mol%; Fig. 5a). K-feldspar is perthitic and contains fine lamellae, is less abundant than plagioclase, and has a wide range of orthoclase contents (Or = 65–90 mol%).

Pseudosection modeling

General remarks

Pseudosection modelings, NCKFMASHT and NCKF-MASHTO, were used to determine the *P–T* conditions recorded by the cordierite-rich domain in sample BTN02A, and by the spinel–garnet-bearing melanocratic layer in sample BTN02F. To calculate the pseudosections, H_2O contents were determined from LOI values, and whole-rock compositions were normalized to a total of 100 wt%. The analytical results of whole-rock compositions are presented in Table S6. Three types of models were calculated 1) a model based on whole-rock composition by XRF disregarding the oxidation state (Model WR), 2) a model based on whole-rock by XRF composition with an assumed oxidation state (Model WR+O), and 3) a model calculated from the composition of the large cordierite with surrounding sillimanite (Fig. 3a) assuming an effective domain composition with different H_2O (Model Crd-domain). The Model WR is based on analytical data, whereas the Model WR+O and the Model Crd-domain involved assumptions to estimate Fe_2O_3 and H_2O. Each model is discussed individually in the following sections.

Model WR - sample BTN02A

We estimated the whole-rock compositions by XRF analysis of cordierite-rich domains from sample BTN02A, and applied them to the modeling of pseudosections (Fig. 8a). Sample BTN02A shows a leucocratic appearance with discontinuous melanocratic layers and lenses cross-cut by leucocratic layers. Based on 6000 point counts from three thin sections, domain contain 29 vol% plagioclase, 28 vol% K-feldspar, 24 vol% cordierite, 8 vol% biotite, 7 vol% sillimanite, 3 vol% garnet, 1 vol % quartz, and 0.2 vol% ilmenite. The inferred peak mineral assemblage for this sample is cordierite–biotite–sillimanite–garnet–plagioclase–melt ± quartz, which is stable on the calculated pseudosection (Fig. 8a) at 6–7 kbar and 780–850 °C. The measured modal proportion of cordierite in this sample (24 vol%) is modelled to occur only at low-*P* and high-*T* conditions of < 5 kbar and 770–850 °C. However, the measured modal abundance of garnet (3 vol%) is modelled to occur at high-*P,* at ~ 6.5–7.1 kbar. The measured modal abundance of sillimanite (7 vol%) is modelled to occur at low-*P* and high-*T* conditions of < 6 kbar, 775–850 °C, and is located outside of the modelled peak *P–T* conditions. The calculated modal abundance of sillimanite increases toward low-*T* and high-*P* conditions, whereas that of garnet increases toward high-*P* and high-*T* conditions (Fig. 8a).
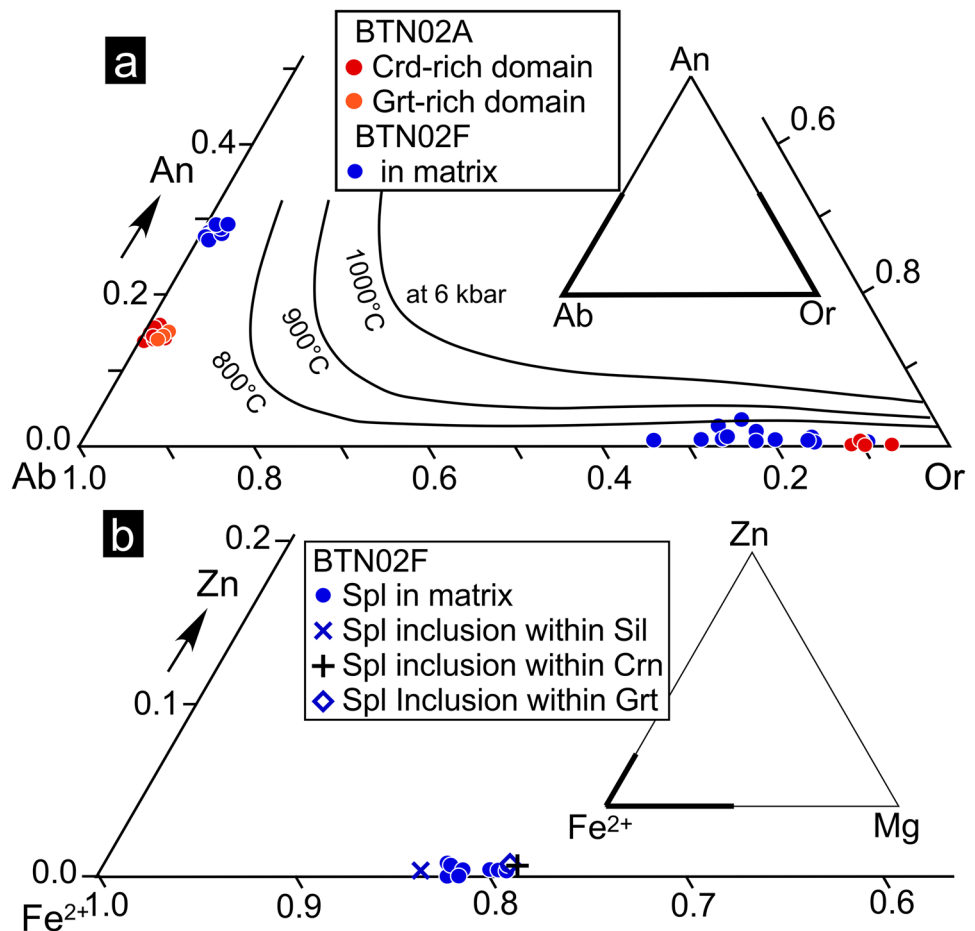
Model WR - sample BTN02F

This sample contains plagioclase (33 vol%), K-feldspar (28 vol%), garnet (17 vol%), spinel (11 vol%), biotite (6 vol%), sillimanite (2 vol%), corundum (1 vol%) and ilmenite (1 vol%). The presence of a small amount of melt is assumable from the leucocratic layer partially invading into the melanocratic layer and occurring as pods (Fig. 2c). The inferred peak mineral assemblage for this sample is garnet–spinel–plagioclase–K-feldspar-ilmenite–melt, which is stable on the modelled pseudosection (Fig. 8b) at 4.5–8.0 kbar and 830–900 °C. The observed mineral assemblage (inferred peak minerals plus biotite–sillimanite–corundum) is stable at 6.2 kbar and 800 °C. The measured modal abundance of spinel (11 vol%) and garnet (17 vol%) suggest low-*P* (< 4.5 kbar) and high-*T* (750–900 °C) conditions based on the positions of the calculated modal proportion isopleths (Fig. 8b). The calculated isopleth for the estimated modal proportion of garnet (17 vol%) is located at ~ 4.5 kbar, where it intersects mineral assemblage characterized by stable cordierite or orthopyroxene, neither of which are present in the assemblage observed in the sample. In

contrast, the measured modal abundance of sillimanite (2 vol%) and corundum (1 vol%) suggest *P–T* conditions of < 6 kbar and 780–800 °C.

Model WR+O - sample BTN02A

We assessed the bulk-rock oxidation state using Fe_2O_3 estimated from average biotite and garnet compositions. A biotite $Fe^{3+}/(Fe^{3+} + Fe^{2+})$ value of 0.08 was applied, based on the typical composition of granulite-facies biotite (Stephenson 1977; Deer et al. 1992), and a garnet $Fe^{3+}/(Fe^{3+} + Fe^{2+})$ value of 0.02, which was estimated from charge balance calculations. The bulk-rock oxidation state was assumed to be Fe_2O_3:FeO = 0.02:2.19, obtained from modal mineral proportions and average mineral Fe^{3+} compositions, according to Palin et al. (2016). In Fig. 9a shows the pseudosection incorporating the estimated Fe_2O_3 content (as "O_2"). The pseudosection topology is near-identical to that calculated for Model WR (Fig. 8a), except for the addition of spinel-stable fields. The inferred peak assemblage is stable on the calculated pseudosection (Fig. 9a) at 6.0–7.2 kbar and 750–850 °C, which is nearly the same as in the WR Model (Fig. 8a).

Fig. 5 **a** Compositional variation of feldspar in terms of An-Ab-Or ternary diagrams. Ternary miscibility gaps are calculated for *P* = 6 kbar using the methods of Fuhrman and Lindsley (1988) and Wen and Nekvasil (1994); **b** Compositional variations of spinel (hercynite) in Fe–Mg–Zn ternary diagrams. Spinel included within sillimanite have lower X_{Mg} than in other textural settings. Spinel in sample BTN02F (Spl-Grt gneiss) have low ZnO contents

Model WR+O - sample BTN02F

The modal proportions of spinel, garnet, and biotite and their average $Fe^{3+}/(Fe^{3+} + Fe^{2+})$ compositions were used to estimate Fe_2O_3 contents of this sample. The $Fe^{3+}/(Fe^{3+} + Fe^{2+})$ ratios for spinel and biotite were estimated to be 0.06 and 0.10, respectively, and the whole-rock Fe_2O_3 is inferred to be 0.57 wt% (0.25 mol%). The Fe^{3+} content of the garnet was found to be negative using an average composition, and based on the charge balance. Figure 9b shows a pseudosection for sample BNT02F incorporating Fe_2O_3 (as O_2). This pseudosection has a different topology to that for the Model WR (Fig. 8b). Spinel stability is restricted to both low-*P* and high-*T* conditions (Fig. 9b), whereas the sillimanite stability field is expanded.

The inferred peak mineral assemblage for this sample is garnet–spinel–plagioclase–K-feldspar–ilmenite–melt, which is stable on modeled pseudosection (Fig. 9b) at < 5.0–7.0 kbar and 800–900 °C. The observed mineral assemblage (inferred peak minerals plus biotite–sillimanite–corundum) is stable at ~ 6.1 ± 0.1 kbar and ~ 800 ± 10 °C (open circle in Fig. 9b). A calculated isopleth of the modal abundance of spinel (11 vol%) is located at 6.1 kbar and 870 °C in the melt–garnet–plagioclase–K-dfeldspar–ilmenite–spinel field (Fig. 9 b) and is assumed to be peak conditions. On the calculated pseudosection (Fig. 9b), sillimanite is stable at 6.1 kbar and < 850 °C, whereas corundum is only stable at 6.1 kbar and < 800 °C. The stability fields of the spinel shift to spinel + sillimanite toward low-*T* conditions. Spinel + sillimanite is further changed to sillimanite + corundum towards low-*T*

condition. The modal abundance of sillimanite (2 vol%) and corundum (1 vol%) are also modeled to stable at 6.1 kbar and 800 °C (open circle in Fig. 9b). The spinels were found in the matrix as inclusions of sillimanite and corundum (Fig. 7a and b). The mineral textural relationships indicate that the spinel was stable first, followed by the growth of sillimanite and corundum. This implies that the rock transversed the IBC *P–T* path from the spinel stability field toward the stability fields of spinel + sillimanite, and subsequently corundum + sillimanite (see arrow in Fig. 9b).

Model Crd-domain - sample BTN02A

This model is only applied to BTN02A to evaluate the texture observed in large cordierite and surrounding sillimanite (Fig. 3a). This rock exhibits a heterogeneous appearance (Fig. 2a), with large cordierite appearing as melanocratic lenses in the cordierite-rich domain. This led us to perform a pseudosection calculation based on the bulk composition of the domain. Domain composition is calculated by assuming sphere cordierite (10 mm in diameter) in a cube matrix (15 mm³). The mineral mode of cordierite (16 vol%) and matrix (84 vol%) was calculated. The mineral mode of the matrix was estimated using point counts, except for cordierite, in large cordierite-bearing thin sections (Fig. 3a).

The estimated domain contains 31 vol% plagioclase, 32 vol% K-feldspar, 16 vol% cordierite, 4 vol% biotite, 16 vol% sillimanite, 1 vol% quartz, and 1 vol% ilmenite. Average mineral compositions by EPMA and modal proportions of minerals are applied to estimate domain composition

Fig. 6 Plane-polarized transmitted-light photomicrographs of garnet-rich domain in sample BTN02A (cordierite-felsic gneiss). **a** Garnet (Grt) porphyroblast is partially replaced by biotite (Bt) and cordierite (Crd); **b** Euhedral cordierite surrounded by biotite in the matrix; **c** cordierite is completely and partially surrounded by garnet. The garnet margin was not replaced by cordierite; **d** Cordierite surrounded by the irregular shaped garnet which shows a sharp crystal margin. Qz = quartz
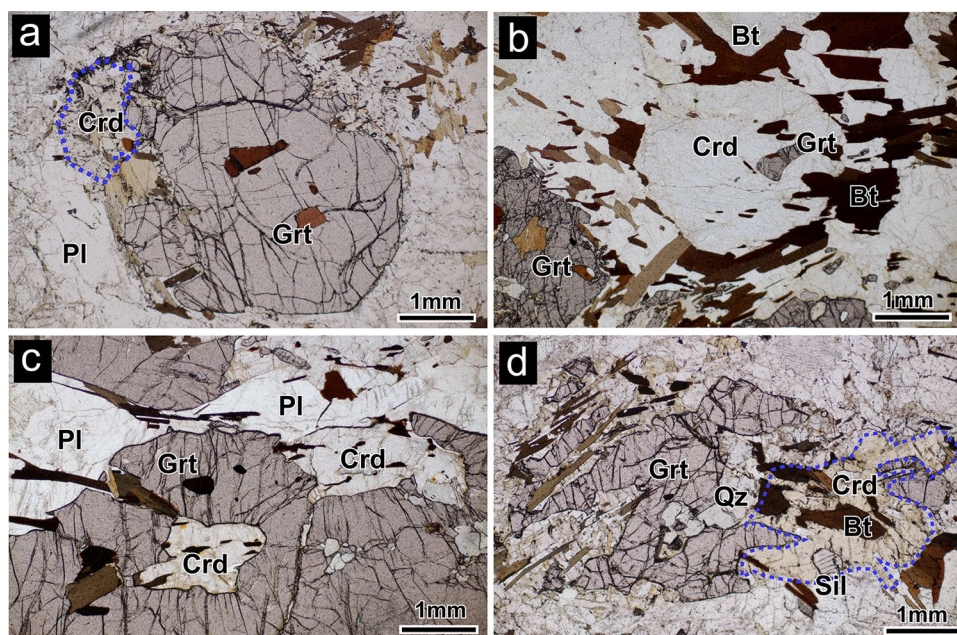
Fig. 7 Plane-polarized transmitted-light photomicrographs showing mineral relationships in sample BTN02F (spinel-garnet gneiss). **a** Garnet porphyroblast and spinel (Spl) in a matrix of plagioclase (Pl). Closeup view of spinel enclosed by thin garnet; **b** Corundum (Crn) and sillimanite (Sil) occur in the matrix; **c** Sillimanite contains micro-inclusions of spinel and corundum and is surrounded by biotite; **d** Corundum contains spinel inclusions
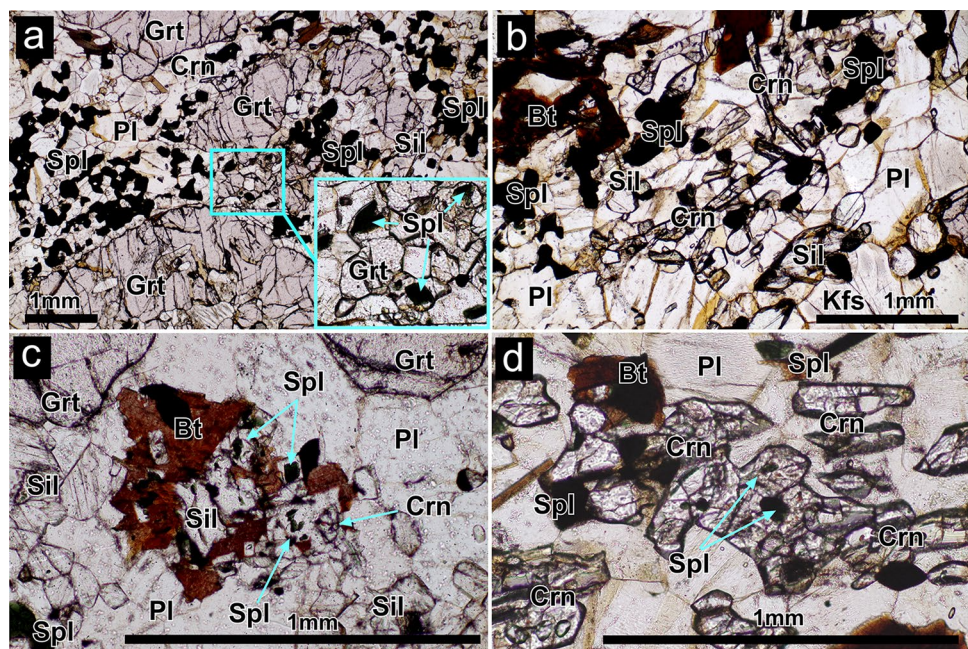
according to Palin et al. (2016). The H_2O contents were inferred from the LOI values because the estimated H_2O from modal proportions of mineral was low (0.28 wt%), and the preliminary calculation gave unrealistic pseudosection results involving orthopyroxene and a wide spinel stability field. Two models of all LOI as H_2O (1.0 LOI) and half LOI as H_2O (0.5 LOI) were applied. Figure 10 shows the results of the two models. The inferred peak mineral assemblage for this domain is biotite–cordierite–plagioclase–K-feldspar–sillimanite–melt, which is stable on the calculated pseudosection at 4.0–6.5 kbar and 720–840 °C (Model H_2O = 1.0 LOI, Fig. 10a) and at 5.8–6.2 kbar and 780–840 °C (Model H_2O = 0.5 LOI, Fig. 10b). The calculated modal abundance isopleths of cordierite (16 vol%) and sillimanite (16 vol%) suggest low-*P* (4.9–5.2 kbar) and high-*T* (830–840 °C) in both models (stars in Fig. 10a and b) are located outside of stability field of inferred peak assemblage. The inferred *P–T* condition of ~ 5.2 kbar and ~ 840 °C estimated from isopleth in Model H_2O = 0.5 LOI is located in spinel–corundum ± sapphirine stable fields, which is inconsistent with our petrographic observation. The inferred *P–T* condition of ~ 4.9 kbar and ~ 830 °C, respectively, estimated from isopleths in Model H_2O = 1.0 LOI is located at the melt–biotite–cordierite–plagioclase–K-feldspar–sillimanite–rutile stable field. Rutile was not present in the measured domain, but was almost the same as the inferred peak assemblage. Therefore, this model is considered more reliable. Euhedral to subhedral sillimanite is present in the matrix and the cordierite margin (Fig. 3a), and the cordierite does not contain sillimanite inclusions. These sillimanites are interpreted to have crystallized in a late phase compared

to cordierite. As the modal abundance of sillimanite in the Model H_2O = 1.0 LOI increases toward high-*P* and low-*T*, sillimanites may crystallize during cooling, crossing the isopleths (gray arrow in Fig. 10a).

Discussion

Metamorphic *P–T* path

Sample BTN02A

Previous studies have described cordierite from Sinnan Rock and Cape Ryugu, which are located within the amphibolite-facies zone in the eastern part of the LHC (Shiraishi et al. 1989). Secondary cordierite has been identified in rocks along the southern part of the Soya Coast at Rundvågshetta, occurring as decompression-related corona textures consisting of cordierite + orthopyroxene and cordierite + sapphirine (Ishikawa et al. 1994; Motoyoshi and Ishikawa 1997; Yoshimura et al. 2008). However, no studies have reported large porphyroblasts of cordierite, and the sample described here is presumed to contain the largest cordierite grains known in the LHC.

Cordierite occurs as variably shaped, subhedral to irregular grains within the different textural domains of sample BTN02A. In cordierite-rich domains, large cordierite porphyroblasts contain biotite inclusions and are in contact with biotite within the matrix (Fig. 6b). The biotite inclusions within cordierite and biotite in the matrix have high TiO_2 contents (~ 6.1 wt%), suggesting crystallization

Model WR
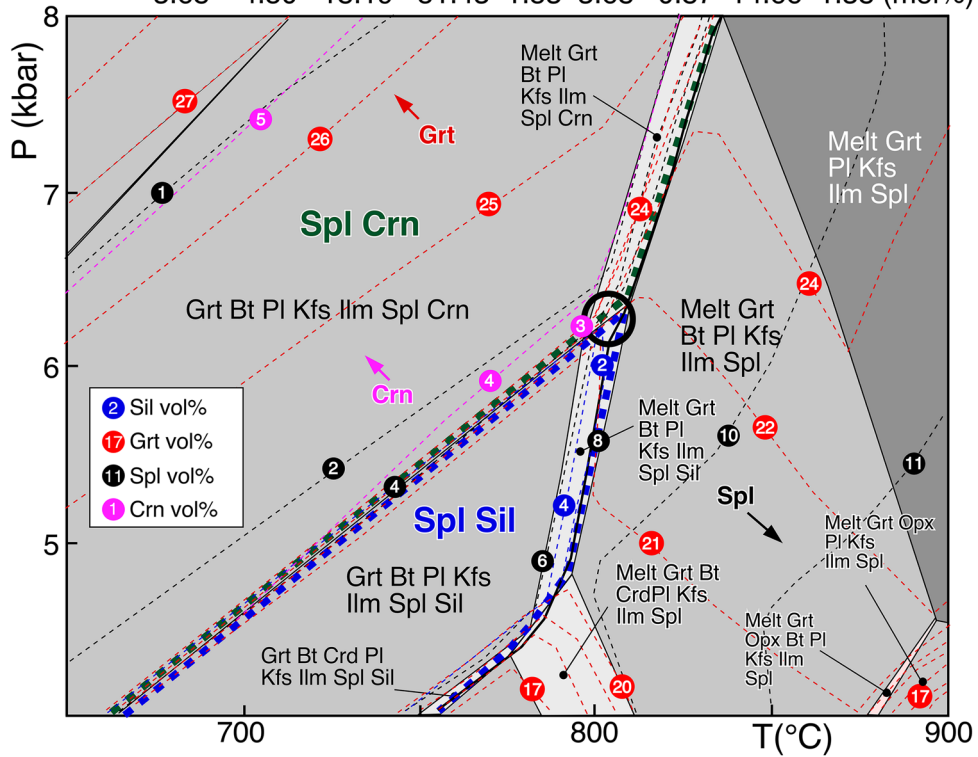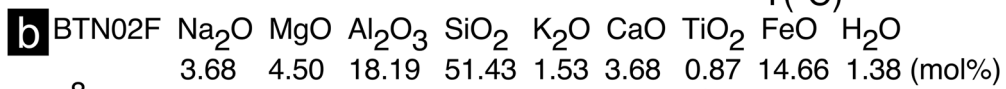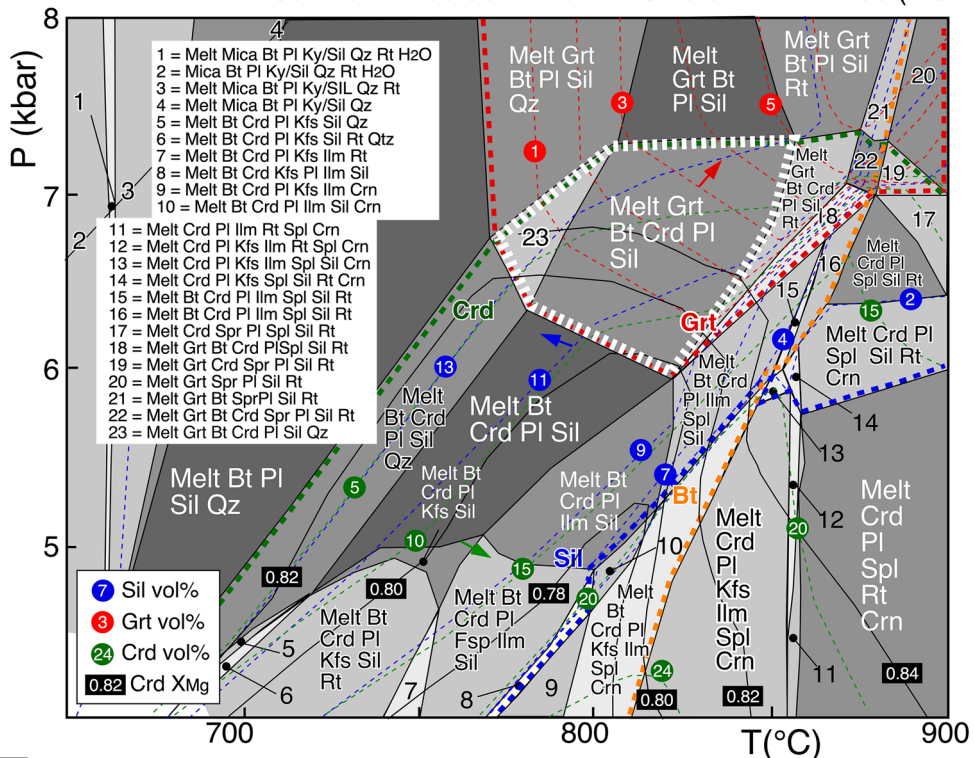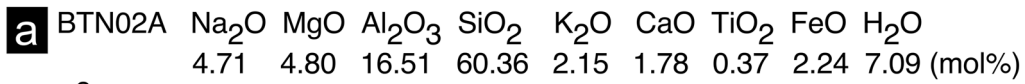
◀**Fig. 8** NCKFMASHT *P–T* pseudosection using the whole-rock composition (Model WR). **a** Result of the cordierite-rich domain of BTN02A. Modal abundance isopleths are shown for garnet, sillimanite, cordierite, and isopleths of [X_{Mg}=Mg/(Mg+Fe)] cordierite are shown. Small arrows represent an increased trend of a modal abundance of each mineral with the same color as isopleths. The white dashed area represents a stable field of the mineral assemblage of melt-garnet-biotite-cordierite-plagioclase-sillimanite±quartz. **b** Result of the of BTN02F. Modal abundance isopleths are shown for garnet, sillimanite spinel and corundum. Open circle represents condition of the mineral assemblage for melt-garnet-biotite-spinel-corundum-sillimanite-plagioclase-K-feldspar-ilmenite. Bt=biotite, Crn=corundum, Grt=garnet, Ilm=ilmenite, Kfs=K-feldspar, Ky=kyanite, Pl=plagioclase, Qz=quartz, Rt=rutile, Spl=spinel, Spr=sapphirine

under high-*T* conditions. Considering the estimated modal proportion (24 vol%) and X_{Mg} values (~0.81) of cordierite and the position of these isopleths in Model WR and Model WR+O, this phase was likely stable at the peak-*T* conditions of ~4.5 kbar and ~825 °C. The estimated conditions located in the spinel and corundum stable fields were inconsistent with those of sample BTN02A, which did not contain either spinel or corundum. Otherwise, inferred pressure values using modal proportion of garnet (3 vol%) suggests 6.5–7.1 kbar but the modal proportion is low. In addition, the estimated peak *P–T* fields (white dashed area in Fig. 8a and b) lacked K-feldspar. Therefore, the estimated conditions are uncertain.

In the case of Model Crd domain, the modal abundance isopleths of cordierite and sillimanite indicate 5 kbar and 830 °C in the H_2O=1.0 LOI model, which is more appropriate than the 0.5 LOI model. In the 0.5 LOI model, spinel, corundum, and sapphire are stable at the inferred *P–T* condition using a modal abundance (Fig. 10b). Sillimanite is typically a late phase, as it is not commonly found as inclusions in large cordierite. Euhedral to subhedral sillimanite and biotite crystallization on the margins of the large cordierite may have progressed during cooling, which is consistent with the increased modal abundance of sillimanite and decreased cordierite (Fig. 10a). The late crystallization of sillimanite occurred within the melt–biotite–cordierite–plagioclase–K-feldspar–sillimanite stability field during cooling after the peak conditions (gray arrow in Fig. 10a). As isopleths of a modal abundance of sillimanite and cordierite have low-angle positive d*T*/d*P* slope, a slight increase in pressure during IBC is likely.

One of the main lines of evidence for an IBC *P–T* path for the Botnnuten gneiss is the growth of euhedral to subhedral sillimanite and biotite at the margins of cordierite in the sample BTN02A (Fig. 3a). The cordierite contains high-TiO_2 biotite and lacks sillimanite inclusions. Similar textural relationships, in which cordierite is overgrown by fine-grained biotite+sillimanite±ilmenite, has been reported

in granulite-facies metapelite from the Rayner Complex, and has been interpreted to reflect a near-IBC path or a counter-clockwise *P–T* trajectory (MacRobertson Land and Northern Prince Charles Mountains; Clarke et al. 1989; Boger and White 2003; Halpin et al. 2007).

Sample BTN02F

In the spinel-garnet-bearing gneiss (sample BTN02F), spinel is of consistent size and is abundant throughout the matrix. As the spinel does not contain significant ZnO, it is unlikely to be a product of the breakdown of staurolite; thus, the spinel-forming reaction remains unclear. Sillimanite and corundum occur together in the matrix with spinel, and garnet locally occurs as overgrowths on spinel rims (Fig. 7a). Sillimanite and corundum often contain spinel and coexist with biotite in the matrix (Fig. 7c), suggesting they crystallized after spinel together with biotite. Phase relations obtained from Model WR+O incorporating O_2 (Fig. 9b) show a different topology than the Model WR ignoring O_2 (Fig. 8b). It implies that the oxidation state is an essential factor in elucidating the stability field of spinel. Boger et al. (2012) calculated a pseudosection considering Fe_2O_3 for spinel-bearing granulites, and their results showed the same phase relations as obtained in the present study, with spinel being stable under low-*P* and high-*T* conditions, and sillimanite under high-*P* and low-*T* conditions; therefore, we consider our *P–T* estimate using Model WR+O to be reliable. The consumption of spinel to form corundum, sillimanite, and biotite is consistent with an IBC path from ~6.1 kbar and ~870 °C toward the garnet–corundum–sillimanite–spinel–biotite–K-feldspar–ilmenite–melt stablility field (at ~6.2 kbar and 800–810 °C) as shown in Model WR+O (Fig. 9b).

Comparison with previous works

Motoyoshi and Shiraishi (1985) used phase relations and geothermobarometry to infer metamorphic *P–T* conditions of 5–6 kbar and 750–800 °C for samples collected from Botnnuten by the First Japanese Antarctic Research Expedition in 1957. Their geothermobarometric results (dashed gray square in Fig. 9a) are consistent with the results of our modeling, and they likely represent the conditions during peak metamorphism. Although these *P–T* conditions are similar to those of the Yamato–Belgica Complex (metamorphic age of 539 Ma: Shiraishi et al. 2003), located ~250 km southwest of the LHC (Fig. 1b), Motoyoshi and Shiraishi (1985) concluded that Botnnuten is a low-pressure part of the LHC based on the occurrence of sillimanite, and geothermobarometric results (garnet–biotite, two-pyroxene and garnet-orhopyroxene geothermometries

Model WR+O
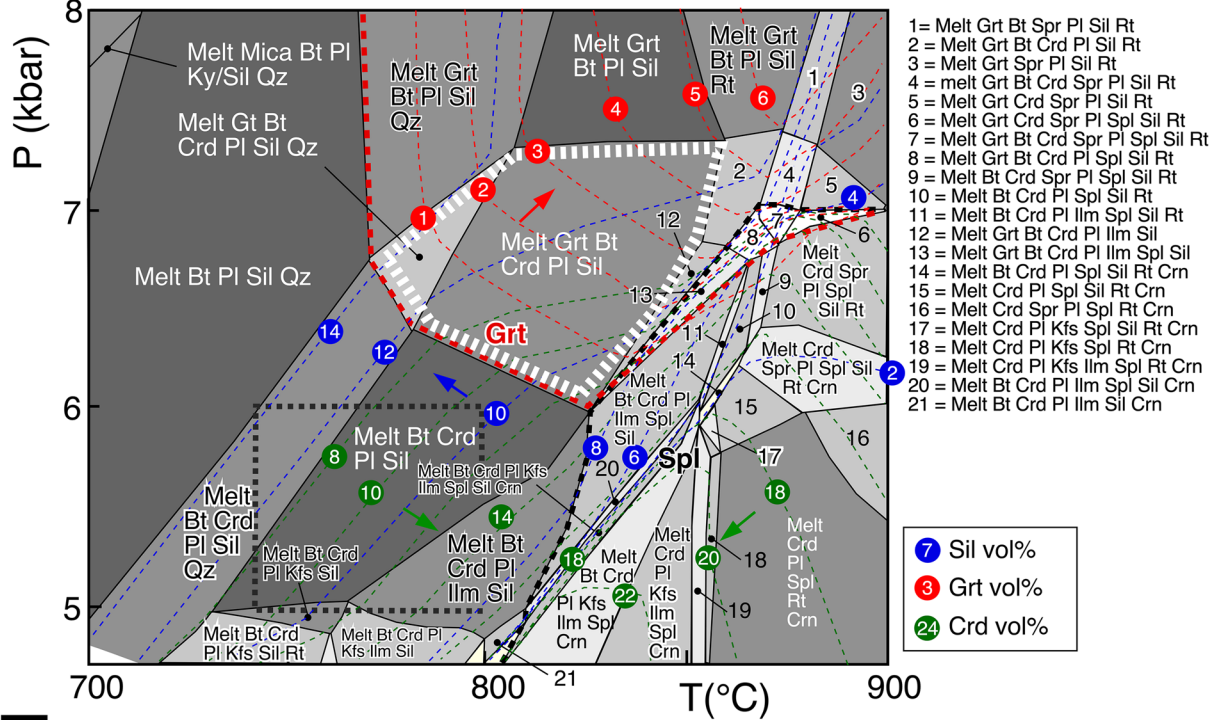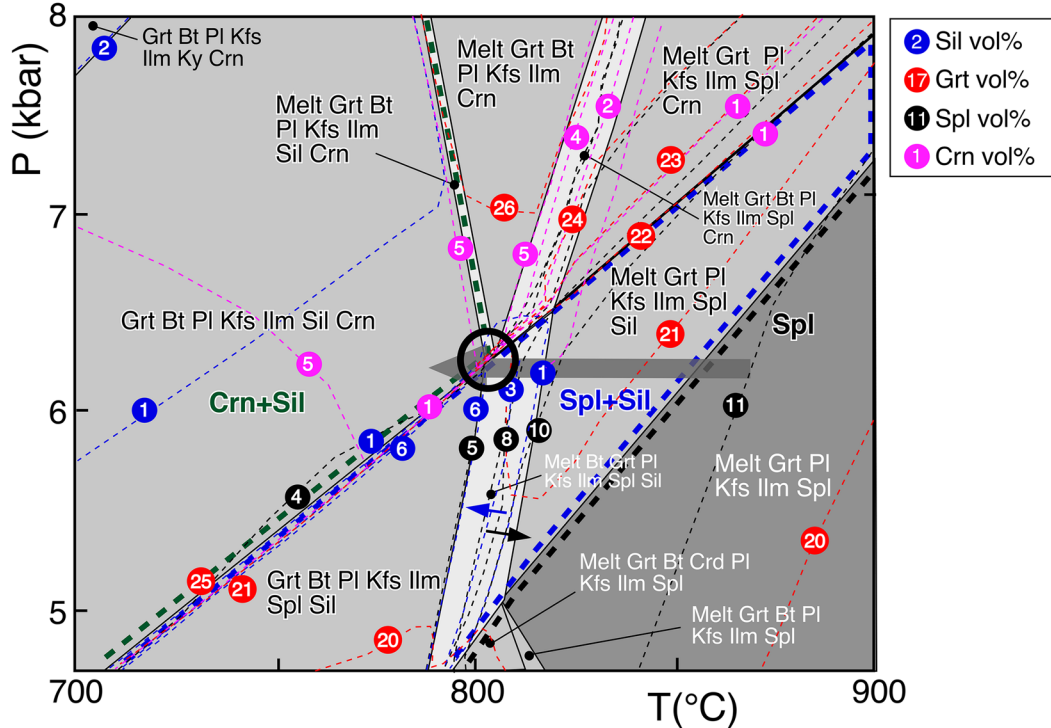
◀**Fig. 9** NCKFMASHTO *P–T* pseudosection using the whole-rock composition (Model WR+O). **a** Result of the cordierite-rich domain of BTN02A. Modal abundance isopleths are shown for garnet, sillimanite and cordierite. The white dashed area represents a stable field of the mineral assemblage of melt-garnet-biotite-cordierite-plagioclase-sillimanite ± quartz. The dashed square represents the *P–T* condition for Botnnuten estimated by Motoyoshi and Shiraishi (1985). **b** Result of the of BTN02F. Modal abundance isopleths are shown for garnet, sillimanite spinel and corundum. Open circle represents condition of the mineral assemblage of melt-garnet-biotite-spinel-corundum-sillimanite-plagioclase-K-feldspar-ilmenite. The arrow indicates the inferred *P–T* path from the point modal abundance isopleth spinel (11 vol%) to the open circle

and garnet–plagiolase–orthopyroxene/clinopyroxene–quartz geobarometries). Despite the absence of staurolite, they assumed a staurolite break-down reaction and a prograde metamorphic history based on the occurrence of spinel.

The ITD-related orthopyroxene–plagioclase symplectite textures that characterize the other parts of LHC have not been identified in the mafic gneisses at Botnnuten. Previous studies of the *P–T* conditions at Rundvågshetta, which is part of the high-grade section of the LHC, yielded maximum pressures of 13–15 kbar and ~900–1000 °C, during the early stages of metamorphism at ca. 580–520 Ma (Kawasaki et al. 2011; Hiroi et al. 2019; Dunkley et al. 2020), which are associated with an ITD path during the retrograde stage. The estimated pressure conditions are much higher than those estimated for Botnnuten in the present study. However, these UHT granulites occur as blocks and thin layers that are restricted to a narrow zone 250–500 m wide in north Rundvågshetta (Motoyoshi and Ishikawa 1997; Yoshimura et al. 2008; Hiroi et al. 2019). Tsunogae et al. (2014) estimated the metamorphic conditions of orthopyroxene-bearing orthogneiss exposed widely in south Rundvågshetta to be < 870 °C, < 9 kbar, and pointed out that the UHT is a local high-temperature event. Thus, the presence of a thickened crust along the entire LHT due to collision is questionable.
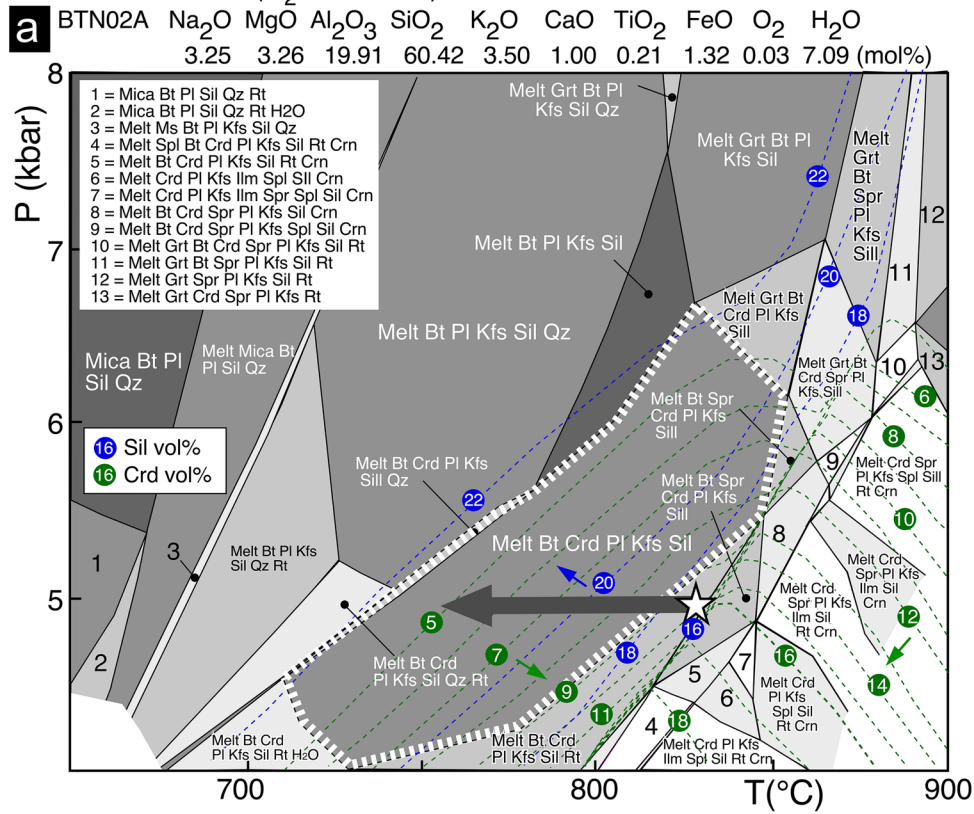
Tectonic significance

In East Antarctica, a counter-clockwise *P–T* path and IBC have been reported from basement rocks of the Rayner Complex (Fig. 1a) in McRobertson Land and the Northern Prince Charles Mountains (Fig. 1a; Boger and White 2003; Halpin et al. 2007; Morrissey et al. 2015). This metamorphism is thought to have been related to the emplacement of voluminous charnockite rocks at 990–900 Ma. In East Antarctica, a counter-clockwise *P–T* path with IBC has also been proposed for the SW terrane of the Sør Rondane Mountains in Eastern Dronning Maud Land (Baba et al. 2013). The SW terrane reached granulite-facies conditions of up to ~9 kbar and ~900 °C (Baba et al. 2013) as a result of tectonic loading
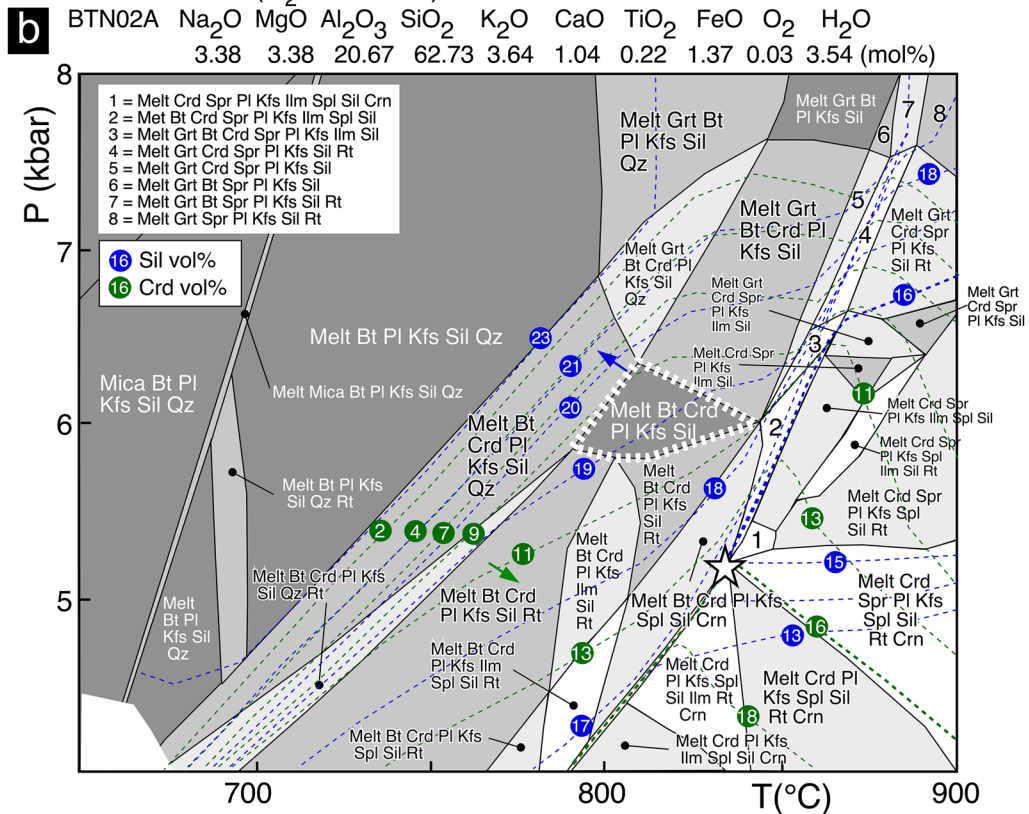
from over-thrusting of the NE terrane (Osanai et al. 2013). The timing of peak metamorphism has been constrained at 650–600 Ma (Osanai et al. 2013). Recently, a counter-clockwise *P–T* path at 620–530 Ma (Kitano et al. 2023) was also proposed for Tenmondai Rock in the LHC (Fig. 1b) based on the textural relation of which the inclusion phase, spinel and ilmenite within sillimanite and kyanite, and kyanite and rutile in the matrix (Baba et al. 2023). However, the tectonic process that caused such metamorphism remains unclear.

The Botnnuten gneisses did not reach high-*P* (~6 kbar), and the geothermal gradient is assumed to be > 40 °C/km (> 1400 °C/GPa). This value is relatively high compared to other instance of granulite–UHT metamorphism at 600–500 Ma (Brown 2007; Brown and Johnson 2019). The high geothermal gradient at Botnnuten is obtained at the peak metamorphism, which is not associated with the ITD and is an unusual result in the continent-continent collisional setting of the LHC. Kelsey and Hand (2015) reviewed tectonic settings to achieve the condition of the high-*T* to UHT granulite facies conditions. They summarized that granulites with near-IBC along counter-clockwise *P–T* paths have historically been interpreted to develop in regions that include synchronous magmatism that either underplates or 'overaccretes' (Wells 1980; Bohlen 1991; Kelsey and Hand 2015 references therein). In the case of the granulite characterized by a near IBC *P–T* path, it is considered to have formed by lithospheric extension tectonics and is deemed to occur at the base of regular thickness crust (Sandiford and Powell 1986; Harley 1989). They also associated with long crustal residence times and low erosion rates (Sandiford and Powell 1986; Kelsey and Hand 2015 further reference therein). The 2D petrological–thermomechanical numerical models by Sizova et al. (2014) predicted that metamorphic rocks recording counter-clockwise metamorphic *P–T* paths followed by IBC could be formed by assuming a hot collision regime, involving shortening and thickening of hot lithosphere (Figs. 3f and 7c in Sizova et al. 2014). Recently, Jiao et al. (2023) summarized the tectonic settings that generate UHT (high-*T* granulite) metamorphism and divided it into 1) an arc–back-arc system involving a thinned lithosphere or orogenic plateaus and 2) an orogenic plateau accompanied by radiogenic heating during thickening. According to their calculated *P–T* paths using numerical modeling for various tectonic settings, the lithospheric peeling model involving mantle upwelling is applicable to the low-*P* and IBC paths inferred in this study (Fig. 5a and h in Jiao et al. 2023). To achieve low-*P* and high-*T* conditions at Botnnuten, the rocks, therefore, likely experienced additional heat flux from the mantle or heat advection due to magmatism at mid-crustal levels, which may cause by crustal extension or crustal peeling (Jiao et al. 2023). In addition, the IBC path recorded

Model Crd-domain (H₂O=1.0 LOI)

a BTN02A Na₂O MgO Al₂O₃ SiO₂ K₂O CaO TiO₂ FeO O₂ H₂O
3.25 3.26 19.91 60.42 3.50 1.00 0.21 1.32 0.03 7.09 (mol%)

Model Crd-domain (H₂O=0.5 LOI)

b BTN02A Na₂O MgO Al₂O₃ SiO₂ K₂O CaO TiO₂ FeO O₂ H₂O
3.38 3.38 20.67 62.73 3.64 1.04 0.22 1.37 0.03 3.54 (mol%)

◀**Fig. 10** NCKFMASHTO *P–T* pseudosection using the large-cordierite domain composition of BTN02A (Model Crd-domain). **a** Result of the Crd-domain assuming $H_2O = 1.0$ LOI. **b** Result of the Crd-domain assuming $H_2O = 0.5$ LOI. Modal abundance isopleths are shown for sillimanite and cordierite. The white dashed area represents a stable field of the mineral assemblage of melt-biotite-cordierite-plagioclase-K-feldspar-sillimanite. Small arrows represent an increased trend of modal abundance of sillimanite and cordierite. A star marks the *P–T* condition estimated from modal abundance isopleths of cordierite and sillimanite. The arrow in **a** indicates the inferred *P–T* path based on the increase and decrease trend of a modal abundance of cordierite and sillimanite

by rocks at Botnnuten implies that they cooled over a long duration at mid-crustal levels due to minimal exhumation of over-thickened crust. These facts conflict with the concept that Botnnuten is a part of the LHC that is considered to have been metamorphosed in an overall orogenic setting.

The past heat flux and advection from the mantle and magmatism would produce voluminous igneous rock; where is the geological evidence preserved? Given that Botnnuten is an isolated inland nunatak, surrounded by an ice sheet, it is difficult to establish its tectonic relationship to surrounding lithologies. In the LHC, Strandnibba and Inhovde (Fig. 1b) have large exposures near Botnnuten; unfortunately, metamorphic *P–T* paths have not been reported in either area. Rocks at Inhovde are hornblende-biotite gneiss, orthopyroxene-bearing orthogneiss (charnockite), garnet-biotite gneiss, mafic granulite, and gneissose granite. The hornblende-biotite gneiss yielded a protolith ages of ca. 981 Ma and a metamorphic age of ca. 574 Ma (Tsunogae et al. 2015). The region from Strandnibba to southern Rundvågshetta is the predominantly orthopyroxene-bearing orthogneiss, yielding protolith ages of 2518–2490 Ma and metamorphic ages of 620–520 Ma (Tsunogae et al. 2014; Dunkley et al. 2020). These regions have similar age relationships with the Botnnuten rocks; thus, Dunkley et al. (2020) regarded Botnnuten a part of the Rundvågshetta Suite (Fig. 1b). However, as mentioned above, the metamorphic conditions differ between the two regions. Overall, no geological evidence suggests voluminous magmatism and additional heat flux from the mantle in nearby exposures to Botnnuten in the LHC during metamorphism at Botnnuten (ca. 590–550 Ma).

The basement rocks of the Yamato Mountains (Fig. 1b) are composed of granulite- to amphibolite-facies metamorphic rocks and voluminous granitoids. The estimated *P–T* conditions of granulite-facies metamorphism in the area are < 6 kbar and ~750 °C (Asami and Shiraishi 1985). Asami and Shiraishi (1985) reported the development of garnet coronas and quartz between anorthite and wollastonite in calc-silicate gneisses from this region, indicating high-temperature isobaric cooling (e.g., Fitzsimons and Harley 1994; Dasgupta and Pal 2001). The granitoids in the Yamato Mountains, which are mainly syenite, occur as lit-par-lit injection layers within the host gneiss (Asami and Shiraishi 1985). If these intrusions were the heat source low-*P*/high-*T* metamorphism would have been likely. Shiraishi et al. (2003) reported two thermal events for Yamato Mountain– an older event (ca. ~620 Ma) thought to be associated with granulite-facies metamorphism and granite magmatism, and a younger event (~535 Ma), which involved amphibolite-facies metamorphism and quartz-monzonite magmatism, which followed the widespread intrusion of syenite pluton. The metamorphic age of 590–550 Ma for the Botnnuten rocks (Dunkley et al. 2020) is slightly older than the timing of widespread syenite magmatism. Therefore, a direct tectonic and metamorphic connection between Botnnuten and the Yamato Mountains remains problematic. To further explore the metamorphic *P–T* path, geochemical and geochronological studies to clarify the magmatism and tectonic setting of the entire LHC, including unexplored localities, are necessary to understand the geological processes, including crustal thinning by extensional or crustal delamination in a continent-continent collisional setting.

Conclusions

We investigated the *P–T* evolution of a suite of cordierite- and spinel-bearing gneisses from Botnnuten, East Antarctica, using mineral chemistry, petrography, and pseudosection modeling, and determined that the rocks at Botnnuten underwent high-*T*/low-*P* metamorphism (~5.0–6.2 kbar and ~850 ± 20 °C) and IBC towards ~800–810 °C. The studied samples are the first cordierite-bearing gneisses to have been identified at Botnnuten, and the first megacrystic cordierite to be observed in the LHC region. Pseudosection models were constructed using 1) whole-rock compositional data (WR), 2) WR incorporating O_2 (WR+O), and 3) a domain composition for the megacrystic cordierite and surrounding minerals. The Botnnuten rocks experienced a high thermal gradient (> 40 °C/km), implying that metamorphism was driven by additional heat flux from the mantle or heating by magmatism. The IBC trajectory suggests that Botnnuten was situated at mid-crustal levels for a long duration, and that the rocks were not rapidly exhumed due to orogenic collapse. This geological process occurred in a part of the orogenic system and was related to the formation of the Gondwana supercontinent. The inferred *P–T* path and peak metamorphic conditions at Botnnuten are similar to those reported for the Yamato Mountains, located approximately 150 km to the southwest of the study area. There is a slight age difference between the metamorphism at Botnnuten and magmatism in the Yamato Mountains, requiring further detailed studies into the connection between the two regions.

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