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(Ultra)high-Pressure Metamorphism in Sedimentary Versus Basaltic Rocks: Insights From a Variscan Mélange in the South Carpathians

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ABSTRACT

Reconstructing the pressure–temperature–time (P – T – t) evolution of (ultra)high-pressure rocks is essential for constraining subduction and collisional dynamics. Subduction zones are often investigated using basic eclogites, whereas nonbasic high-pressure rocks remain comparatively understudied, potentially hindering a comprehensive understanding of complex subduction systems. The Bughea Complex is a modern-style mélange within the pre-Alpine basement of the South Carpathians (Romania) that hosts a wide variety of eclogitic rocks, providing an opportunity to investigate their response to subduction processes. Here, we integrate phase equilibrium modelling, rutile thermometry, zircon and apatite U–Pb dating coupled with trace element analysis and garnet Lu–Hf dating to reconstruct the P – T – t evolution of three different eclogitic rocks from the Bughea Complex: a kyanite eclogite, a basic eclogite and an eclogitic schist. The kyanite eclogite is interpreted to derive from a continental sedimentary protolith based on its continental crust-like trace-element patterns and detrital-like zircon U–Pb age spectrum (ranging from Archean to Cambrian). It records the highest peak pressure conditions (~ 3.2 – 3.6 GPa at 700°C – 820°C), consistent with ultrahigh-pressure metamorphism, followed by decompression and cooling to < 1.6 GPa and $< 675^{\circ}\text{C}$. In contrast, the basic eclogite, interpreted to derive from altered oceanic crust based on its MORB-like trace-element patterns, and the eclogitic schist, which exhibits trace-element signatures similar to trench sediments in volcanic-arc settings, record lower peak pressures (~ 2.0 – 2.3 GPa at 580°C – 610°C and ~ 1.9 – 2.4 GPa at 530°C – 600°C , respectively). The former records near-isothermal decompression (~ 1.3 GPa at 585°C – 615°C), whereas the latter records decompression and minimal heating (~ 1.0 – 1.2 GPa at 600°C – 615°C). Distinct garnet Lu–Hf ages (469 ± 96 to 345 ± 21 Ma) suggest diachronous metamorphism during a prolonged Variscan subduction, whereas apatite U–Pb age (ca. 340 Ma) likely records late-stage cooling and exhumation. Differences in P – T – t records may reflect mélange rheology and variations in subduction dynamics, whereas limited heating during retrogression and compositional buoyancy appear critical for preserving peak conditions, particularly in nonbasic rocks. Overall, these findings demonstrate that continental-derived metasedimentary rocks can preserve evidence of extreme pressure conditions not always recorded by associated basic eclogites, highlighting the importance of integrating nonbasic high-pressure rocks into P – T – t reconstructions of subduction-related mélanges to achieve a more complete understanding of complex collisional systems.

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1 | Introduction

Eclogites can provide direct evidence of deep subduction and collisional processes, including mechanisms and rates of ocean-basin closure, crustal differentiation, fluid–rock interaction, recycling of elements and the exhumation of deeply buried rocks (Rubatto and Hermann 2001; Rapp et al. 2003; Spandler and Pirard 2013; Warren 2013; Tsujimori and Ernst 2014; Brown and Johnson 2019). Typical eclogites have basic composition, which favours the development of diagnostic parageneses (Carswell 1990). These eclogites are often derived from subducted oceanic crust and are considered indicators of suture zones, making them a central focus of studies aimed at unravelling deep lithospheric processes (e.g., Davis and Whitney 2006; Groppo et al. 2009; Omrani et al. 2013; Laurent et al. 2018). However, ongoing discoveries of coesite and diamond in continental rocks have demonstrated that a wide range of nonbasic rocks can also record information on deep subduction processes (Chopin 2003; Gilotti 2013; Ducea 2016). This has sparked debate on how such extreme pressure conditions are achieved and recorded and which rock types are most suitable for recording it (Agard et al. 2009; Erdman and Lee 2014; Chapman et al. 2019). Notably, nonbasic high-pressure rocks have received limited attention, despite their potential to record extreme pressure conditions, including those derived from sedimentary protoliths (e.g., Reinecke 1991, 1998; Negulescu et al. 2009; Manzotti et al. 2022, 2025).

Here, we investigate high-pressure rocks with eclogitic parageneses derived from sedimentary and igneous protoliths to reconstruct their origins and P – T – t trajectories during the pre-Alpine evolution of the South Carpathians. We integrate petrography, mineral chemistry, whole-rock geochemistry, phase equilibrium modelling, Zr-in-rutile thermometry, zircon and apatite U–Pb dating coupled with trace element analyses and garnet Lu–Hf dating to reconstruct the metamorphic evolution of these rocks and evaluate their tectonic significance and possible factors controlling the potential of different subducted lithologies to record extreme pressure conditions.

2 | Geological Background

The Carpathians are part of the Alpine-Himalayan Orogenic System, forming a semiannular orocline spanning across central and eastern Europe (Figure 1a). Their southern branch (namely, the South Carpathians) forms an E–W-trending mountain chain extending through central Romania (Figure 1b).

The South Carpathians was built during the collision between the Dacia microplates, the East European Craton and the Moesian Platform, along with the closure of the Vardar Ocean around 115–80 Ma, during the Alpine Orogeny (Săndulescu 1994; Csontos and Vörös 2004). The pre-Alpine basement of the South Carpathians corresponds to remnants of the Dacia microplates, consisting of two main tectonostratigraphic units (Iancu and Seghedi 2017 and references therein): (i) the upper Getic and Supragetic Units, which comprise a stack of asymmetric, fan-like nappes composed of Palaeozoic igneous and low- to high-grade metamorphic rocks and (ii) the lower Danubian Unit, composed of Proterozoic–Palaeozoic

medium-grade metamorphic rocks intruded by Palaeozoic granitoids. These units are separated by Jurassic ophiolitic complexes, representing remnants of a former ocean compressed during the Alpine collision (Figure 1c).

High-pressure metamorphic rocks are widespread across the Getic Unit, including eclogitic rocks derived from basic, ultrabasic and sedimentary protoliths, recording contrasting metamorphic histories. In the central part of the South Carpathians (namely, the Lotru Metamorphic Suite), basic eclogites occur as relatively small pods hosted by medium-grade micaschists and gneisses (Iancu et al. 1998; Medaris Jr. et al. 2003; Săbău and Massonne 2003). Whole-rock-garnet Sm–Nd dating of an eclogite and a biotite-garnet amphibolite from this area indicates high-pressure metamorphism around 360–340 Ma (Medaris Jr. et al. 2003), associated to the Variscan Orogeny. This orogenic event resulted from the amalgamation of northern Gondwana and southern Laurussia around 360–340 Ma (Franke et al. 2021; Schulmann et al. 2025 and references therein). Similar age ranges have also been reported for eclogites across the European Variscides, including those in basement inliers within Alpine orogens (e.g., Stosch and Lugmair 1990; Schmädicke et al. 1995; Giacomini et al. 2007; Bröcker et al. 2009; Herwartz et al. 2011; Lotout et al. 2018; Trapp et al. 2021; Vanardois et al. 2022; de Hoÿm de Marien et al. 2023; Tual et al. 2025).

In the easternmost part of the South Carpathians (namely, the Leaota Massif), eclogites are found within a thin, coherent unit known as the Bughea Complex, interpreted as a modern-style *mélange* (Iancu et al. 1998; Săbău 2000; Negulescu and Săbău 2015) (Figure 1d). The Leaota Massif is formed by distinct tectonostratigraphic units, including the Bughea Complex, which are separated by major Variscan shear zones developed around 360–335 Ma (Axente et al. 2008).

In the Bughea Complex, eclogites are hosted by semipelitic (e.g., micaceous gneiss, amphibole-garnet-bearing schist and chloritoid-bearing micaschist) and basic rocks (e.g., garnet-bearing amphibolite). These eclogites are subdivided into the following four types: common eclogite, Cr-rich eclogite, metagabbro-norite and very high-pressure eclogite (Săbău 2000; Negulescu and Săbău 2015 and references therein).

The ‘common’ eclogites represent the most abundant eclogite type in the Bughea Complex. They display geochemical signatures of mid-ocean ridge basalts (MORB) and contain metamorphic assemblages ranging from omphacite + garnet $\pm$ quartz to more complex assemblages including phengite, amphibole, kyanite and/or zoisite. Reported peak P – T estimates for these eclogites include ~2.0 GPa at 780°C–860°C (Maruntiu et al. 1997), ~1.2–2.7 GPa at 520°C–650°C (Săbău 2000) and ~1.5–2.2 GPa at 580°C–745°C (Medaris Jr. et al. 2003). Phengite and amphibole Ar–Ar ages around 355–340 Ma obtained from a retrogressed eclogite were interpreted as the timing of cooling (Axente et al. 2008).

The Cr-rich eclogite shows a geochemical signature similar to picrites and consists of nodules formed by Cr-rich staurolite, kyanite, amphibole, mica and chromite blebs, embedded in a high-pressure matrix formed by clinopyroxene, garnet, amphibole, zoisite and rutile. The metagabbro-norite contains relics

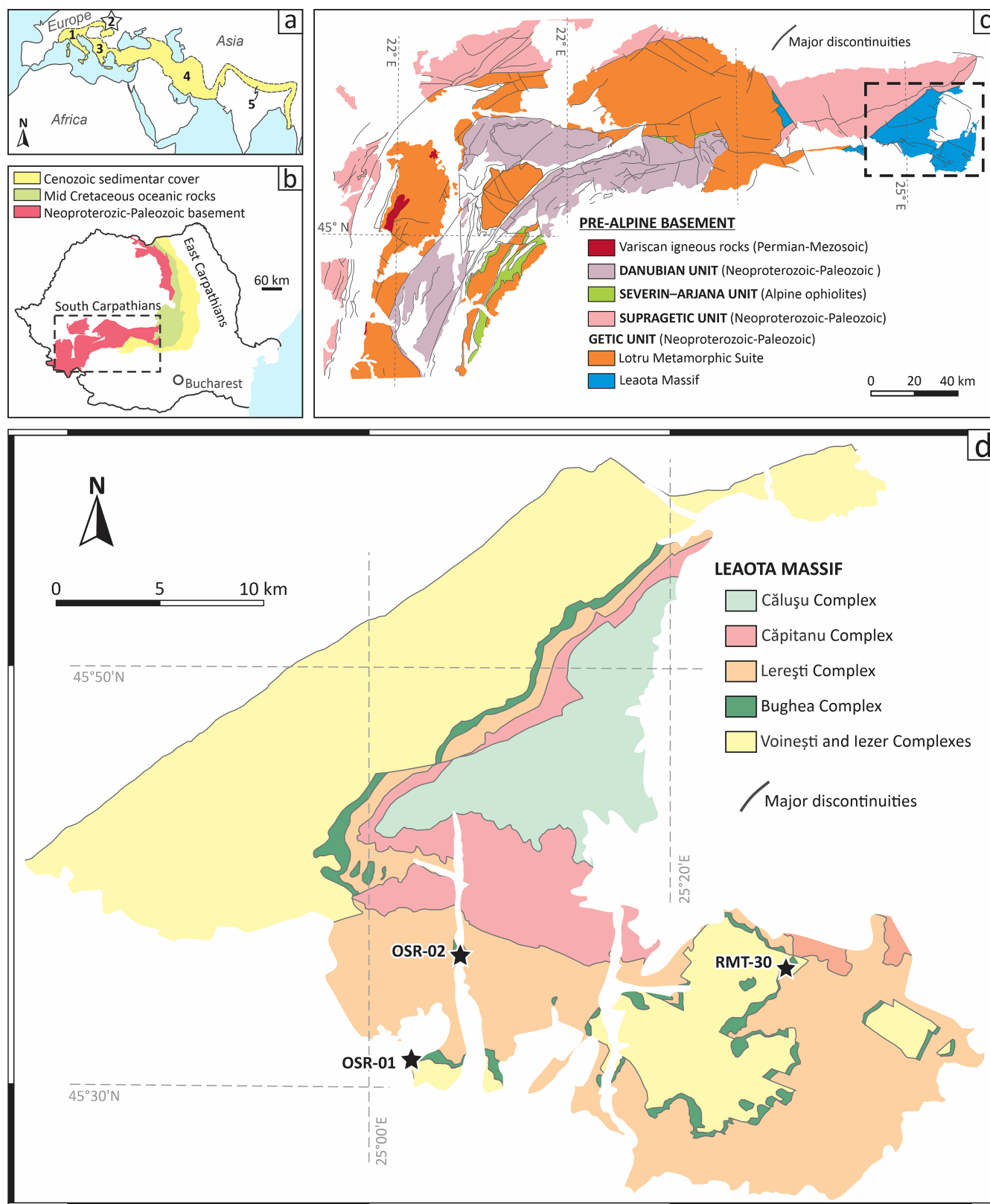


FIGURE 1 | Geological framework of the South Carpathians. **(a)** Extent of the Alpine–Himalayan Orogenic System in Europe, Middle East and Asia (redrawn after Furnes et al. 2020). Numbers refer to the mountain chains: 1 – Alps; 2 – Carpathians; 3 – Dinarides–Hellenides; 4 – Afro–Arabian collisional zone; 5 – Himalayas. **(b)** Simplified tectonic map of the Carpathians in the Romanian territory (simplified after Săndulescu 1994). **(c)** Geological map of the pre–Mesozoic basement of the South Carpathians (redrawn after Săbău and Negulescu 2012). **(d)** Geological map of the Leaota Massif, in the easternmost South Carpathians (redrawn after Negulescu et al. 2018). Black stars represent the location where the eclogites were sampled.

of igneous assemblages (augite, enstatite, labradorite, biotite, ilmenite, spinel and Mg-hornblende) overprinted by metamorphic jadeite-rich clinopyroxene, garnet and Si-rich phengite (Negulescu and Săbău 2015).

The very high-pressure eclogite (named by Negulescu et al. 2009 according to the definition of Liou et al. 1998) is interpreted to have originated from a sedimentary protolith based on the bulk-rock geochemistry (Negulescu et al. 2009) and contains garnet porphyroblasts, jadeite-rich clinopyroxene, kyanite, quartz, amphibole, phengite and paragonite (Săbău 2000). Reported P – T estimates using multiequilibrium thermobarometry indicate a clockwise path with peak pressure conditions of ~ 3.0 GPa at 780°C – 840°C (Săbău 2000). A zircon U–Pb age of ca. 550 Ma, obtained for this eclogite, was interpreted as the timing of ultrahigh-pressure metamorphism in the Bughea Complex (Negulescu 2006).

3 | Methods

In this study, we selected three eclogitic rocks from the Bughea Complex for petrological and geochronological analyses: a kyanite eclogite (#RMT-30), a kyanite-bearing amphibole-phengite eclogite (#OSR-01B) and an eclogitic schist (#OSR-02C) (Figure 1d). We acknowledge that sample #RMT-30 does not exhibit the typical basic composition expected for eclogites *stricto sensu* (see Results and Discussions). However, as demonstrated in this study, it displays mineral assemblage, modal proportions and P – T conditions characteristic of eclogites. Accordingly, we adopt the nomenclature ‘eclogite’ for this rock in a broader sense (*lato sensu*).

3.1 | Whole-Rock Geochemical Analysis

Geochemical data acquisition took place at ALS Geochemistry (Vespasiano, Brazil and Perth, Australia). Weight fractions of major and minor oxides were determined by using inductively coupled plasma atomic emission spectroscopy, while trace elements were analysed via inductively coupled plasma mass spectrometry. The whole-rock geochemical data for the three samples, together with those of other rocks from the Bughea Complex compiled from the literature, are presented in the Supporting Information S1.

3.2 | Mineral Chemical Analysis and Compositional Mapping via Electron Probe Microanalyzer

Mineral compositions were acquired using a JEOL JXA-8200 Electron Probe Microanalyzer (EPMA) at the Institute of Geological Sciences, University of Bern (UNIBE, Switzerland) and a JEOL JXA-8230 EPMA at the Universidade Estadual Paulista (UNESP, Brazil). Spot analyses were conducted on amphibole, biotite, clinopyroxene, zoisite-clinozoisite, garnet, plagioclase, paragonite and phengite using a 1- to $5\text{-}\mu\text{m}$ spot size, an acceleration voltage of 15 keV and a probe current of 20 nA . The counting times on the peak position were set to 20 s for major elements and 30 s for minor elements. The garnet

structural formula was calculated using the approach proposed by Droop (1987). Biotite, clinopyroxene, zoisite-clinozoisite, paragonite, plagioclase and phengite structural formulas were calculated using the MinPlot software (Walters 2022). The amphibole structural formula was calculated using the approach of Li et al. (2020). Chemical classification of amphibole and clinopyroxene is according to Hawthorne et al. (2012) and Morimoto (1988), respectively. All mineral chemical data are presented in Supporting Information S2–S4. Diagrams of mineral chemical discrimination and classification are presented in Supporting Information S5. Mineral abbreviations are according to Warr (2021), but the phengite abbreviation (Ph) is according to Whitney and Evans (2010).

Four areas within the polished thin sections were selected for x-ray mapping. Two areas from #RMT-30 were analysed using the EPMA at the University of Bern: (i) the Map I depicts a garnet porphyroblast, its inclusions and matrix phases; and (ii) the Map II depicts matrix phases. The other two areas, from #OSR-01B and #OSR-02C, were analysed at the Universidade Estadual Paulista. Analytical setups are summarized in Supporting Information S2–S4. The resulting x-ray maps were converted into quantitative compositional maps using XMapTools 4.4 software (Lanari et al. 2014, 2019; Lanari and Tedeschi 2025). In this paper, we only present garnet compositional maps. Additional compositional maps are provided in Supporting Information S6.

In situ trace-element analyses of rutile included in garnet and along the grain boundaries of the main phases were performed on all three samples using EPMA equipment at the Universidade Estadual Paulista. Analytical conditions followed Luvizotto et al. (2009) and Fumes et al. (2024), using a focused beam, an accelerating voltage of 20 keV , a probe current of 100 nA and a spot size of $2\text{ }\mu\text{m}$. The analysed elements include Si, Al, Cr, Sb, Sn, Ta, Fe, Zr, Nb and Ti. Analysis yielding Si contents higher than $300\text{ }\mu\text{g/g}$ were excluded to avoid undetected zircon contamination, following the approach proposed by Zack et al. (2004) and Luvizotto and Zack (2009). Rutile standards R10 and Sy were used for calibration (Luvizotto et al. 2009) and Zr-in-rutile temperatures were calculated using the calibration of Kohn (2020). Additional information on operating conditions, along with rutile chemical data and calculated temperatures, is provided in Supporting Information S7.

3.3 | Phase Equilibrium Modelling

To constrain prograde and peak P – T conditions of the three samples, phase diagrams were calculated using whole-rock and reactive bulk compositions, accounting for garnet chemical fractionation (Marmo et al. 2002; Tinkham and Ghent 2005; Lanari and Engi 2017). The reactive bulk compositions were estimated as follows. First, the proportion of each garnet zone was determined relative to the whole-rock. Mineral phase proportions were estimated from optical observations, whereas the relative proportions of garnet zones were estimated from compositional maps. The volumes of garnet zones were calculated assuming spherical geometry, with the volumes of outer zones obtained by subtracting those of the core and previously formed internal zones (Lanari and Engi 2017). Representative compositions for each garnet zone were selected from EPMA analyses. The oxides

contents in each zone were multiplied by its corresponding proportion to estimate the amount of material sequestered during each growth stage. These values were then subtracted from the measured whole-rock composition to generate stage-specific reactive bulk compositions.

Retrograde conditions were constrained using phase diagrams based on reactive bulk compositions approximated from local bulk compositions (Lanari and Engi 2017; Rodrigues et al. 2026). These were obtained from small, mapped areas showing disequilibrium textures and retrograde assemblages that partially replace peak mineral phases. H_2O and $X_{Fe^{3+}}$ contents were estimated using P - X and T - X phase diagrams. All bulk compositions are presented in Table 1.

Phase diagrams were calculated with Theriak-Domino software (De Capitani and Petrakakis 2010) considering the 11-component MnNCKFMASHTO (MnO - Na_2O - CaO - K_2O - FeO - MgO - Al_2O_3 - SiO_2 - H_2O - TiO_2 - O_2) chemical system and the ds62 dataset of Holland and Powell (2011), with the following activity-composition models: ilmenite-haematite (White et al. 2000), magnetite-spinel (White et al. 2002), feldspars ($C1$ plagioclase model) (Holland and Powell 2003), epidote, olivine (Holland and Powell 2011), biotite, chlorite, white micas, orthopyroxene (White, Powell, Holland, et al. 2014; White, Powell, and Johnson 2014), amphibole, clinopyroxene (Omp model) and melt (Green et al. 2016). Here, only the phase diagrams calculated for peak pressure condition are presented, whereas additional phase diagrams for prograde and retrograde stages are provided in Supporting Information S8–S10.

3.4 | Zircon and Apatite U–Pb–Trace Element Analyses

Zircon and apatite from the kyanite eclogite (#RMT-30) were analysed for U–Pb isotopes and trace element contents via laser ablation quadrupole inductively coupled plasma mass spectrometry at the GeoHistory Facility, Curtin University

(Australia). Zircon grains were concentrated through standard mechanical, gravimetric and magnetic procedures and further embedded in rounded polished epoxy mounts. Cathodoluminescence images of the zircon grains were obtained using a JEOL QUANTAX Compact30 scanning electron microprobe at the Centro de Pesquisa Manoel Teixeira da Costa, Universidade Federal de Minas Gerais (CPMTC-UFMG, Brazil). Apatite was analysed in situ on a polished thin section.

The analysis was carried out using a RESOLUTION SE excimer laser (193-nm wavelength) with a Laurin Technic S155 sample cell coupled to an Agilent 7900 ICP-MS. A ‘squid’ mixing device (Laurin Technic) was used to smooth the aerosol pulses between the laser and mass spectrometer. The sample cell was flushed with ultrahigh purity He (320 mL/min) and N_2 (4 mL/min), both passed through an inline Hg trap. The low flow rate of N_2 was added to the carrier gas before the ICP torch to enhance sensitivity (Hu et al. 2008). The samples and reference materials were ablated using a laser fluence of 2.5 J/cm², repetition rate of 10 Hz (zircon) and 7 Hz (apatite) and a circular beam of 33 μ m (zircon) and 50 μ m (apatite). Each analysis consisted of 30 s of baseline acquisition, followed by 30-s sample ablation and 10-s fast washout. Instrument tuning was performed using NIST SRM 610 glass to optimize plasma conditions, maximize instrument sensitivity and minimize oxide interferences.

Unknowns were bracketed by a suite of zircon, apatite and glass reference materials. Zircon GJ-1 (Jackson et al. 2004) and McClure Mountain apatite (Schoene and Bowring 2006) were employed as primary reference materials for U–Pb isotopes, and NIST SRM 610 and trace element, respectively. Zircon 91500 (Wiedenbeck et al. 2004) and Plešovice (Sláma et al. 2008), Madagascar (MAD-2; Thomson et al. 2012) and FC-Duluth apatite (Schmitz et al. 2003; Thomson et al. 2012), and NIST SRM 612 were employed as secondary reference materials and treated as unknowns. Data reduction was performed in Iolite4 (Paton et al. 2011) using an exponential with linear equation to account for U–Pb downhole fractionation, and the built-in ‘Trace

TABLE 1 | Reactive bulk compositions (wt.% of oxides) used for calculating isochemical P - T phase diagrams.

	SiO ₂	TiO ₂	Al ₂ O ₃	FeO _T	MgO	CaO	MnO	Na ₂ O	K ₂ O	H ₂ O	X _{Fe³⁺}
Kyanite eclogite (#RMT-30)											
M₁ (prograde)	54.98	0.90	21.22	9.01	3.23	4.00	0.05	5.68	0.92	0.80	0.15
M₂ (prograde)	55.14	0.91	21.23	8.81	3.21	3.99	0.04	5.74	0.93	0.56	0.15
M₃ (peak)	57.49	1.03	21.21	6.16	2.79	3.67	0.00	6.57	1.06	0.41	0.05
M₄ (retrograde)	51.53	2.22	25.95	4.31	3.25	4.75	0.04	7.12	0.82	1.50	0.05
Kyanite-bearing amphibole–phengite eclogite (#OSR-01B)											
M₁ (prograde)	49.22	1.46	16.42	11.67	8.76	7.83	0.17	3.42	1.05	0.65	0.25
M₂ (peak)	50.69	1.63	15.65	10.09	8.86	7.88	0.14	3.87	1.19	0.65	0.25
M₃ (retrograde)	52.81	1.15	12.46	5.78	7.80	9.54	0.05	5.25	0.65	1.50	0.30
Eclogitic schist (#OSR-02C)											
M₂ (peak)	57.63	0.44	16.01	7.44	5.63	9.15	0.12	2.47	1.11	0.85	0.01
M₃ (retrograde)	57.67	0.08	12.88	1.82	8.75	12.02	0.02	5.09	1.00	1.50	0.01

Element' data reduction scheme using ^{18}Si and ^{43}Ca in stoichiometric proportion as internal calibration for zircon and apatite, respectively. The U–Pb diagrams and dates were calculated using IsoplotR (Vermeesch 2018), with dates and uncertainties being presented at the two-standard error (2SE). Isotopic and trace element data, including those from reference materials employed for zircon and apatite analysis, are presented in Supporting Information S11 and S12. Glass reference material yields trace element data within 5%–10% compared to the reference values.

3.5 | In Situ Garnet Lu–Hf Analysis

In situ isotopic analyses of garnet were conducted in all three samples via triple quadrupole laser ablation inductively coupled plasma mass spectrometry at the GeoHistory Facility, Curtin University (Australia). The analytical setup for garnet Lu–Hf analysis includes a Coherent LR 193-nm ArF excimer laser with a Laurin Technic S155 sample cell coupled to an Agilent 8900 ICP-MS/MS using NH_3/He gas in the reaction cell (20% NH_3 in He premixed gas) following Ribeiro et al. (2024). This configuration enables measurement of Lu and Hf isotopes after Simpson et al. (2021). A 'squid' mixing device (Laurin Technic) was used to smooth the aerosol pulses between the laser and mass spectrometer. The sample cell was flushed with ultrahigh purity He (320 mL/min) and N_2 (4 mL/min), both passed through an inline Hg trap. A low flow rate of N_2 was added to the carrier gas before the ICP torch to enhance sensitivity (Hu et al. 2008). Garnet samples and reference materials were ablated using a fluence of $\sim 3.5\text{ J/cm}^2$, repetition rate of 10 Hz and a circular laser beam of $193\text{ }\mu\text{m}$. Garnet reference materials (GWA-1 and GWA-2; Ribeiro et al. 2024) and NIST SRM 610 were ablated using a circular laser beam of 130 and $33\text{ }\mu\text{m}$, respectively. Each analysis consisted of 30 s of baseline acquisition, followed by 30-s sample ablation and 10-s fast washout. Instrument tuning was performed using NIST SRM 610 glass in single quadrupole (no gas mode) to optimize plasma conditions, minimize oxide interferences and maximize ^{175}Lu sensitivity using a circular laser beam of $50\text{ }\mu\text{m}$. After shifting to MS/MS mode, the ICP-MS was tuned to maximize sensitivity for Hf reaction products using NH_3/He premix reaction gas. To check for potential systematic errors in Lu/Hf from pulse/analogue (P/A) conversion in the 8900-detector system, we performed repeated measurements of the NIST SRM 610 using different spot sizes and detector modes for ^{175}Lu (33 and $64\text{ }\mu\text{m}$), resulting in a P/A factor of ~ 0.18 .

The Lu–Hf isochron dates were calculated using the inverse isochron regression (Li and Vermeesch 2021) in IsoplotR (Vermeesch 2018) with ^{176}Lu decay constant after Söderlund et al. (2004). Inverse isochron dates and analytical uncertainties are stated at 2SE (95% confidence). Garnet GWA-1 (1267.0 ± 3.0 ; Ribeiro et al. 2024) was employed as reference material to account for instrumental and matrix effects following Ribeiro et al. (2024), and GWA-2 ($934.7 \pm 1.4\text{ Ma}$; Ribeiro et al. 2024) was employed as a secondary age check. The GWA-1 yielded a matrix-uncorrected inverse isochron age of $1267 \pm 19\text{ Ma}$ ($n=30$, MSWD=0.9), indicating no offset compared to the isotope dilution Lu–Hf reference age. The GWA-2 yielded a

matrix-uncorrected inverse isochron age of $938 \pm 30\text{ Ma}$ ($n=29$, MSWD=0.9) in agreement with the reference age. All garnet Lu–Hf data are presented in Supporting Information S13.

4 | Results

4.1 | Sample Description and Mineral Chemical Data

4.1.1 | Kyanite Eclogite (#RMT-30)

Sample #RMT-30 is a fine- to coarse-grained rock containing porphyroblasts of garnet ($\sim 25\text{ vol.}\%$) embedded within a decussate matrix formed by clinopyroxene ($\sim 36\%$), amphibole ($\sim 12\%$), albite ($\sim 9\%$), white mica (both phengite and paragonite, $\sim 7\text{ vol.}\%$) and kyanite ($\sim 5\%$) (Figure 2a–c). Quartz ($\sim 4\%$), pyrite ($\sim 2\%$), apatite, zoisite, rutile and zircon (all $< 1\%$) are accessory phases.

Anhedral to subhedral poikiloblastic garnet occurs as highly fractured porphyroblasts (around 0.5- to 1.5-cm diameter), often mantled by composed amphibole and albite coronas exhibiting local magnetite or clinozoisite intergrowth. Amphibole and albite also occur filling intragranular microfractures (Figure 2c,d). The mapped garnet porphyroblast exhibits a smooth concentric chemical zoning with the following four main zones (Figures 3a–c and S5.1c): (i) the core (Grt_1 : $\text{Alm}_{61-62}\text{Prp}_{19-21}\text{Grs}_{15-18}\text{Sps}_{1.5-2.5}$); (ii) the inner mantle (Grt_2 : $\text{Alm}_{57-59}\text{Prp}_{25-26}\text{Grs}_{14-16}\text{Sps}_{0.5-1.5}$); (iii) the outer mantle (Grt_3 : $\text{Alm}_{56-58}\text{Prp}_{23-25}\text{Grs}_{15.5-18}\text{Sps}_{0.3-0.4}$); and (iv) rims (Grt_4 : $\text{Alm}_{57-58}\text{Prp}_{25-27}\text{Grs}_{15-17}\text{Sps}_{0.5-0.6}$). Garnet porphyroblasts exhibit complex textural zoning, depicted by randomly oriented inclusions of clinopyroxene, quartz, paragonite, kyanite, amphibole, apatite, rutile, pyrite and zoisite (Figure 2e). Zoisite ($< 0.1\text{ mm}$) is present only as scattered inclusions at the garnet core, sometimes in direct contact with paragonite. The inner mantle lacks kyanite inclusions, whereas, in the outer mantle, kyanite is present and paragonite is absent. Amphibole inclusions decrease in abundance outwards and disappear at the rims. All other phases are ubiquitous through the zones. The rims are inclusion-poor.

Anhedral to subhedral clinopyroxene (up to 3.0 mm) is omphacite-jadeite and is broadly replaced by amphibole + plagioclase $\pm$ clinozoisite intergrowths (Figure 2f,g). In the matrix, clinopyroxene displays relatively sharp concentric zoning, characterized by core regions with X_{Mg} ($\text{Mg}/[\text{Mg} + \text{Fe}^{2+}]$) = 0.40–0.60 and X_{Na} ($\text{Na}/[\text{Na} + \text{Ca}]$) = 0.80–0.90 (jadeite), whereas rims show X_{Mg} = 0.65–0.75 and X_{Na} = 0.55–0.70 (omphacite) (Figures S5.1a,b and S6.1g,h). Clinopyroxene inclusions in garnet exhibit patchy zoning, with compositional ranges similar to both core and rims of matrix clinopyroxene (X_{Mg} = 0.32–0.94 and X_{Na} = 0.60–0.92). Anhedral to subhedral kyanite (up to $\sim 3\text{ mm}$) shows undulose extinction and is often replaced by aggregates of white mica, clinozoisite and by amphibole + plagioclase intergrowths (Figure 2h). Quartz, white mica, rutile and zircon are common inclusions in omphacite and kyanite, with quartz and white mica rarely displaying radial microcracks (Figure 2e).

Subhedral to euhedral white mica (up to 2.0 mm) consists of phengite and paragonite, both of which are common phases

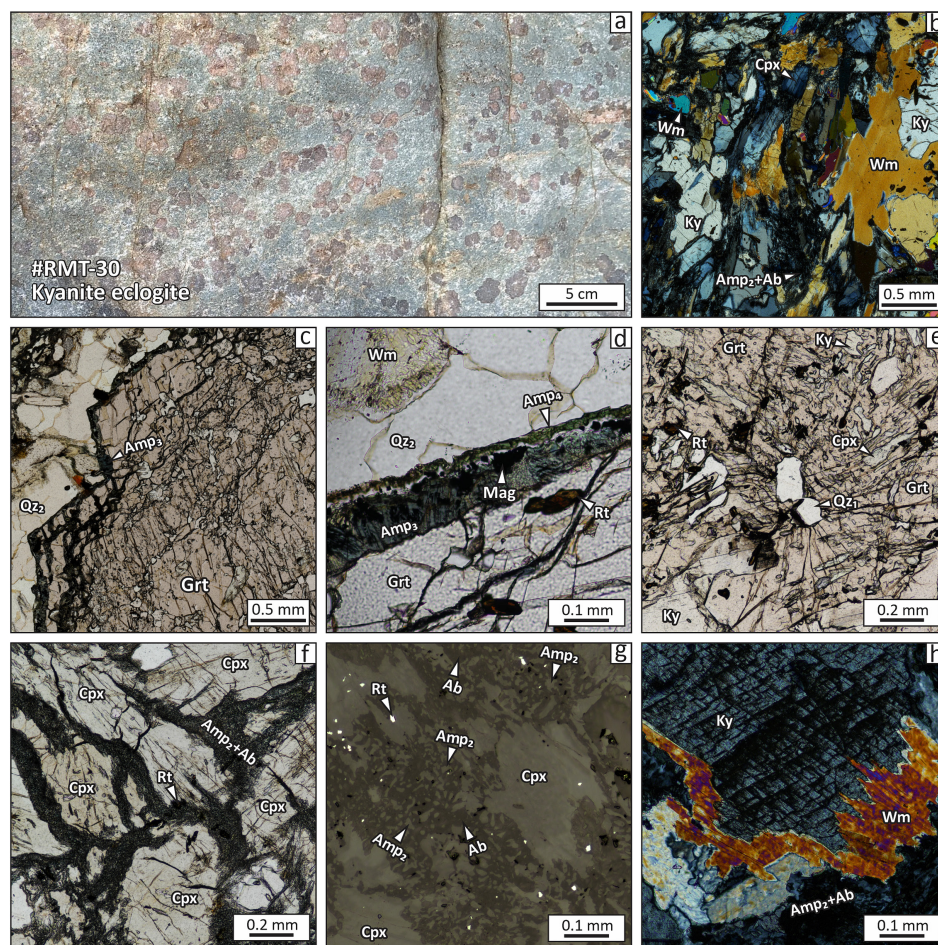


FIGURE 2 | General features of the kyanite eclogite (#RMT-30). **(a)** Macroscopic view of the studied eclogite. **(b–g)** Photomicrographs of the main textural features. **(b)** General aspect of the matrix. **(c)** General aspect of a garnet porphyroblast. **(d)** Coronas of amphibole around garnet, with local intergrowth of magnetite. **(e)** Inclusions of quartz in garnet with radial microcracks. **(f–g)** Symplectites of $\text{Amp}_2 + \text{Ab}$ around clinopyroxene. **(h)** White mica around kyanite.

in the matrix, sometimes occurring as coronas around kyanite (Figure 2h). Phengite is often mantled by amphibole + plagioclase intergrowths and/or replaced by biotite and paragonite. It exhibits patchy zoning in which Si varies between 3.29 and 3.42 cations per formula unit (c.p.f.u.), with lower values occurring close to garnet and kyanite (Figures S5.1e and S6.1i).

The following four types of amphiboles are distinguished: (i) Na–Ca to Na amphibole included in garnet core and mantles (Amp_1 , glaucophane–katophorite; Figure S5.2a,b); (ii) Na–Ca to Ca amphibole forming intergrowths with albite $\pm$ clinozoisite around omphacite–jadeite, kyanite and phengite (Amp_2 , pargasite–sadanagaite–taramite; Figures 2f,g and S5.2b–d); (iii) Na–Ca amphibole inner corona around garnet and along its intragranular microfractures (Amp_3 , taramite; Figures 2c,d and S5.2b); and (iv) Ca amphibole outer corona around garnet (Amp_4 , Fe actinolite; Figures 2d and S5.2d). Plagioclase is albite–oligoclase ($\text{Ab}_{84-99}\text{An}_{1-16}$; Figure S5.1d) and occurs forming intergrowths and composite coronas.

Subhedral to euhedral rutile and zircon (up to ~ 0.35 and ~ 0.1 mm, respectively) occur as rounded to prismatic crystals in the matrix—occasionally within intergrowths—and as inclusions in garnet, clinopyroxene, kyanite, white mica and quartz. Rutile

commonly exhibits ilmenite lamellae. Chemically, rutile exhibits $\text{Fe} = 3077\text{--}8828 \mu\text{g/g}$, $\text{Cr} = 103\text{--}245 \mu\text{g/g}$ and $\text{Nb} = 843\text{--}1020 \mu\text{g/g}$, with $\log (\text{Cr}/\text{Nb})$ ranging from -0.94 to -0.58 , typical of pelitic-felsic protoliths (Meinhold et al. 2008). Rutile included in garnet exhibits Zr contents ranging from 56 to $105 \mu\text{g/g}$, whereas rutile from the matrix shows $59\text{--}82 \mu\text{g/g}$ of Zr (Figure S14.1).

4.1.2 | Kyanite-Bearing Amphibole-Phengite Eclogite (#OSR-01B)

Sample #OSR-01B is a fine- to medium-grained rock, formed by omphacite (~ 43 vol.%), garnet ($\sim 22\%$), phengite ($\sim 18\%$) and amphibole ($\sim 15\%$). Quartz ($\sim 2\%$), kyanite, apatite and rutile (all $< 1\%$) are accessory phases. It shows a well-developed foliation marked by oriented omphacite, phengite, amphibole, quartz and rutile (Figure 4a–c).

Subhedral to euhedral garnet (up to 3.5 mm) occurs as syn- to postkinematic grains that contain nonoriented inclusions of quartz, omphacite, rutile and apatite and display distinct textural zoning characterized by cores rich in quartz, omphacite and amphibole inclusions and inclusion-poor rims (Figure 4a–c). Locally, coronas of amphibole around garnet are observed

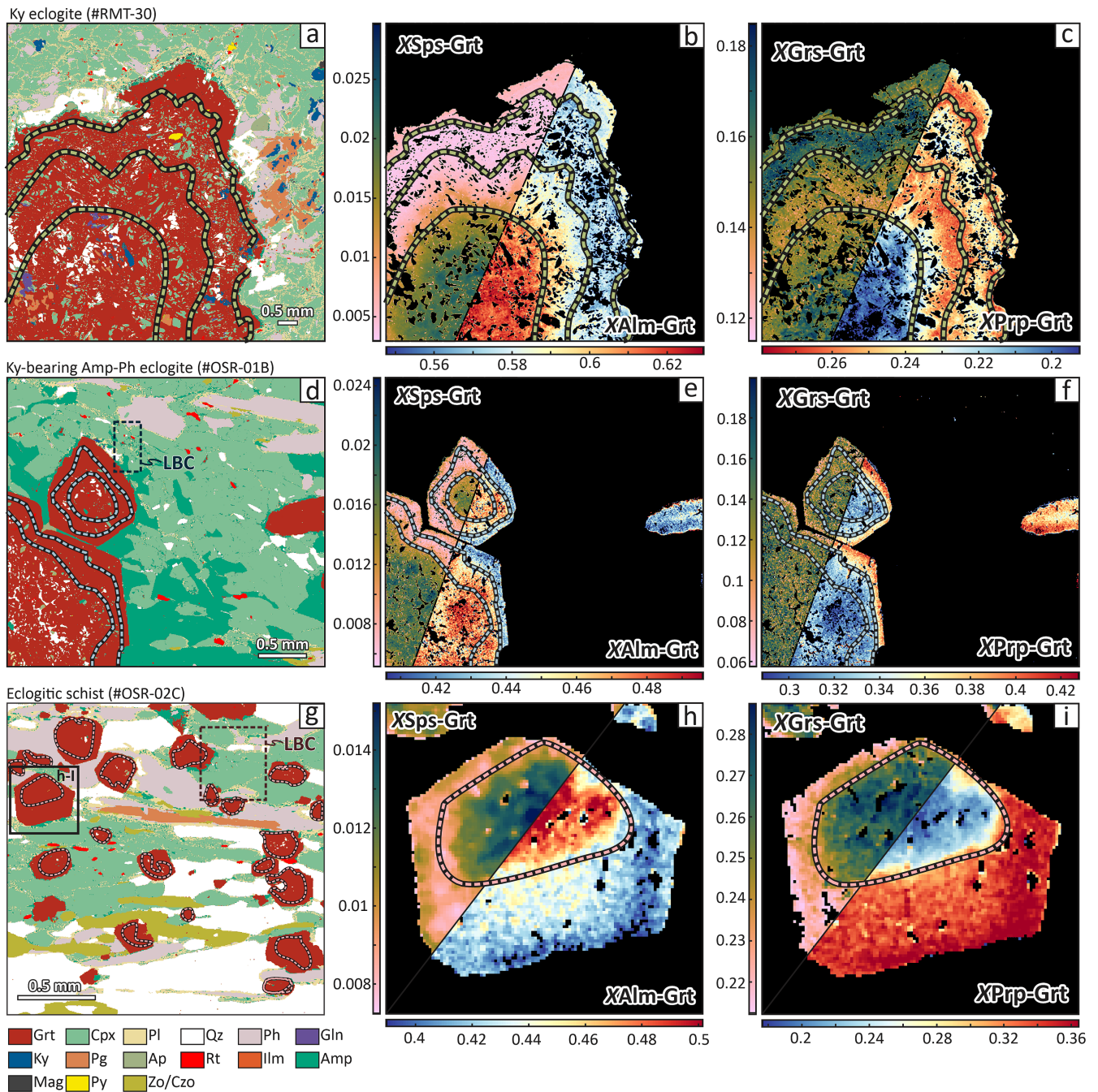


FIGURE 3 | Compositional maps depicting chemical zoning of garnet in the (a–c) kyanite eclogite (#RMT-30), (d–f) kyanite-bearing amphibole–phengite eclogite (#OSR-01B) and (g–i) eclogitic schist (#OSR-02C). Dashed curves represent the approximate boundaries between different garnet chemical zones.

(Figure 4d). The mapped garnet grains exhibit smooth concentric zoning with the following three main zones (Figures 3d–f and S5.1c): the core (Grt₁: Alm_{45–49}Prp_{29–32}Grs_{19–21}Sps_{1.8–2.2}), the mantle (Grt₂: Alm_{45–49}Prp_{30–34}Grs_{19–21}Sps_{0.5–1.5}) and rims (Grt₃: Alm_{42–44}Prp_{37–40}Grs_{16–18}Sps_{0.6–1.0}).

Subhedral omphacite (up to 1.65 mm) hosts inclusions of quartz, rutile and phengite and is locally wrapped by amphibole + plagioclase ± Ca clinopyroxene symplectites (Figure 4e). It is chemically homogeneous, showing $X_{Mg} = 0.78–0.85$ and $X_{Na} = 0.45–0.47$ (Figures S5.1a,b and S6.2f,g). Inclusions of omphacite in garnet crystals display

similar chemical characteristics. Within the symplectites, diopside intergrowths with $X_{Mg} = 0.91–0.93$ and $X_{Na} = 0.16–0.19$ are observed.

Three types of amphiboles are distinguished, including: (i) subhedral Na–Ca amphibole (up to 2.45 mm), which is dominant, oriented and hosts inclusions of phengite, rutile and quartz (Amp₁, winchite–kathophorite–taramite; Figures 4b,c, S5.2b and S6.2i); (ii) Na–Ca to Ca amphibole coronas around garnet (Amp₂, pargasite–sadanagaite–kathophorite; Figures 4d, S5.2b–d and S6.2i); and (iii) thin Ca–amphibole forming symplectites with plagioclase ± Ca clinopyroxene that replace omphacite and

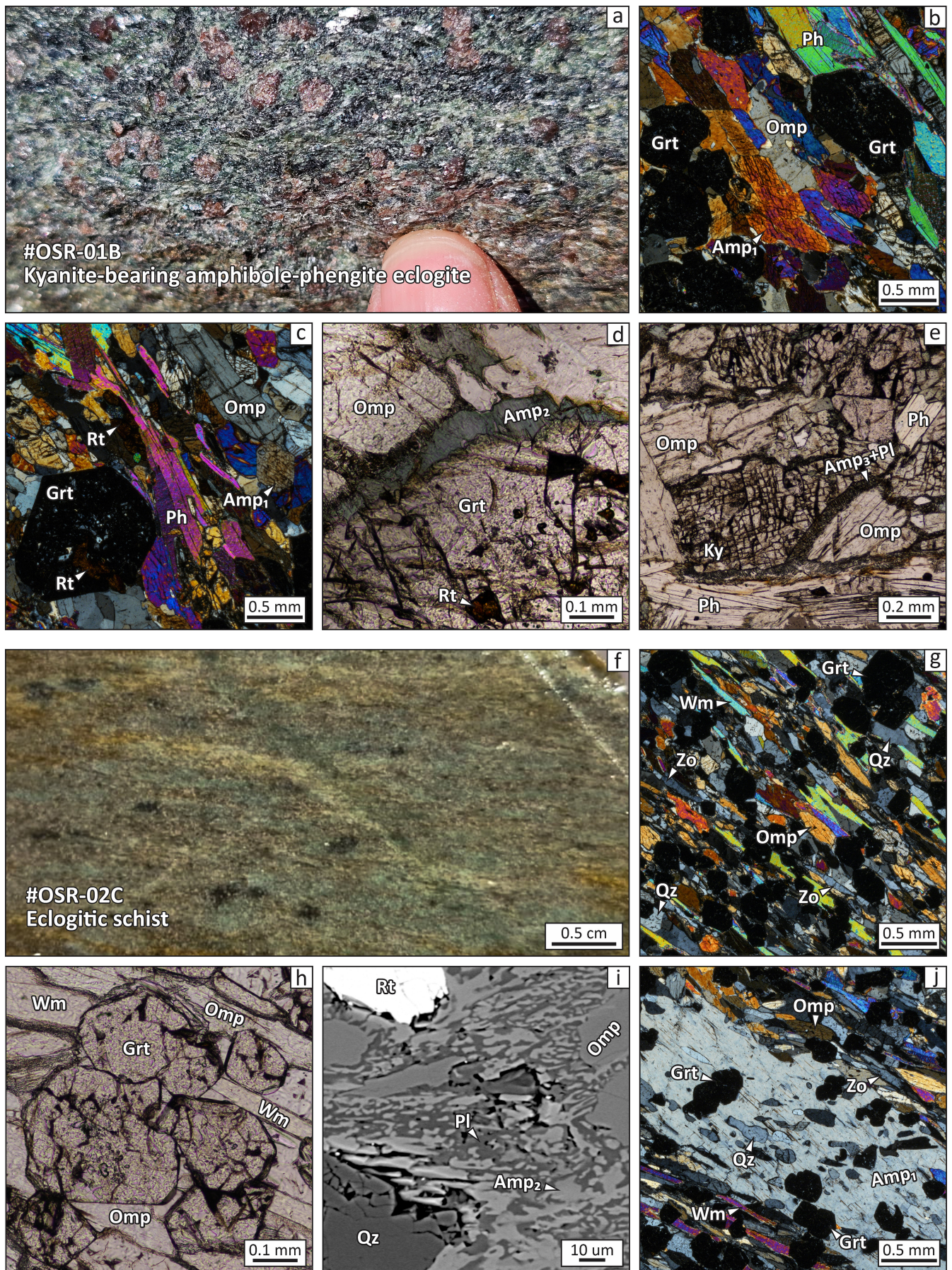


FIGURE 4 | Legend on next page.

FIGURE 4 | General features of the (a–e) kyanite-bearing amphibole–phengite eclogite (#OSR-01B) and (f–j) eclogitic schist (#OSR-02C). (a, f) Macroscopic view of the studied eclogites. (b, c, g) Oriented fabric of both eclogites. (d) Corona of Amp₂ around garnet. (e, i) Amp₃ + Pl symplectites around omphacite. (h) Syn- to post-kinematic garnet. (j) Inclusion-rich amphibole porphyroblast.

Amp₁ (Amp₃, pargasite; Figures 4e, S5.2c and S6.2i). Plagioclase is albite-oligoclase (Ab_{81–90}An_{10–19}; Figure S5.1d) and only occurs in the symplectites.

Subhedral to euhedral phengite (up to 3 mm) contains inclusions of rutile and omphacite, is commonly wrapped by Amp₃ + plagioclase symplectites and is locally replaced by clinozoisite and/or biotite. It displays concentric Si chemical zoning (3.25–3.36 c.p.f.u.), with the highest values occurring at the cores (Figures S5.1e and S6.2h).

Anhedral rutile (up to 0.75 mm) is weakly oriented, ubiquitous in the matrix and as inclusions in the main phases and commonly contains ilmenite lamellae. It exhibits Fe = 3212–8153 µg/g, Cr = 246–860 µg/g and Nb = 44–102 µg/g, with log (Cr/Nb) values ranging from 0.47 to 1.06, consistent with mafic protoliths (Figure S14.1; Meinhold et al. 2008). Zr contents in rutile included in garnet vary between 120 and 170 µg/g, whereas rutile in the matrix exhibits 98–166 µg/g of Zr.

4.1.3 | Eclogitic Schist (#OSR-02C)

Sample #OSR-02C is a fine- to medium-grained rock, composed of quartz (~30 vol.%), omphacite (~22%), garnet (~20%), white mica (phengite and paragonite; ~12%), zoisite (~7%) and amphibole (~5%). Rutile, apatite, sulphides (pyrite and chalcopyrite) and plagioclase (all < 1%) are accessory phases. It shows a well-developed foliation marked by oriented quartz, white mica, clinopyroxene, zoisite, amphibole and rutile (Figure 4f,g).

Subhedral to euhedral garnet (up to 0.6 mm) occurs as nonoriented, inclusion-poor grains, locally replaced by amphibole. It is syn- to postkinematic and hosts scattered nonoriented inclusions of quartz, apatite, rutile and omphacite (Figure 4g,h). The mapped grains show a smooth concentric to patchy zoning, with two main zones (Figures 3g–i and S5.1c): the core (Grt₁: Alm_{46–52}Prp_{18–22}Grs_{25–28}Sps_{1.0–1.5}) and the rims (Grt₂: Alm_{42–45}Prp_{31–35}Grs_{22–24}Sps_{0.8–1.0}).

Subhedral omphacite (up to 1.2 mm) hosts inclusions of quartz and rutile and is often surrounded by amphibole + plagioclase symplectites (Figure 4i). Its X_{Mg} content is mostly homogeneous, varying between 0.90 and 0.95, whereas X_{Na} displays patchy to concentric zoning, with the highest values around 0.40–0.43 and the lowest ranging between 0.38 and 0.41 (Figures S5.1a,b and S6.3f,g). Within the symplectites, diopside intergrowth with X_{Mg} = 0.90 and X_{Na} = 0.16 is observed.

Subhedral to euhedral white mica (up to 1.25 mm) comprises phengite and paragonite, and both are locally wrapped by amphibole + plagioclase symplectites. Phengite is dominant, contains inclusions of omphacite and rutile and exhibits Si = 3.28–3.40 c.p.f.u., which displays a patchy zoning (Figures S5.1e and S6.3h). Locally, muscovite rims with Si = 2.95 c.p.f.u. are observed.

The following three types of amphibole are distinguished: (i) inclusion-rich porphyroblasts (up to 5 mm) that host inclusions of all major phases (Amp₁; Figure 4j); (ii) Ca amphibole forming symplectites with plagioclase (Amp₂, pargasite-tremolite; Figures 4i and S5.2c,d); and (iii) amphibole coronas after garnet (Amp₃). Plagioclase is oligoclase-albite (Ab_{72–94}An_{6–27}; Figure S5.1d) and only occurs in symplectites.

Anhedral rutile (< 0.1 mm) is oriented and occurs both as inclusions in the main phases and in the matrix, locally within the Amp₂ + Pl symplectites and often exhibits ilmenite lamellae. It shows Fe = 728–6435 µg/g, Cr = 613–926 µg/g and Nb = 556–612 µg/g, with log (Cr/Nb) values ranging between 0.02 and 0.22, consistent with mafic protoliths (Figure S14.1; Meinhold et al. 2008). Scattered rutile included in garnet yielded Zr contents of 62–95 µg/g, whereas rutile in the matrix exhibits 50–110 µg/g of Zr.

4.2 | Whole-Rock Geochemical Data

The kyanite eclogite (#RMT-30) shows high SiO₂, Al₂O₃ and Na₂O + K₂O and low CaO and Mg# [= MgO/(MgO + FeO_T)] contents relative to the other eclogites from the Bughea Complex (Figure S15.1). Its chondrite-normalized rare earth elements (REE) profile resembles that of the continental crust with slight relative depletion in LREE and enrichment in heavy REE (HREE) (Figure 5a). In the mantle-normalized trace element diagram, it exhibits a Cs depletion relative to the Post Archean Australian Shale profile (PAAS; Nance and Taylor 1976) and the average upper continental crust (HGUCC; Hu and Gao 2008) and lacks the negative Y anomaly characteristic of the PAAS (Figure 5b).

The kyanite-bearing amphibole-phengite eclogite (#OSR-01B) has a basic composition, similar to the ‘common’ eclogites from the Bughea Complex, with relatively low SiO₂ and Al₂O₃ and high TiO₂, CaO and Mg# (Figure S15.1). Its chondrite-normalized REE profile broadly resembles that of MORB with relative enrichment in LREE and MREE (Figure 5a). In the mantle-normalized trace element diagram, this sample deviates from typical basaltic crust compositions mostly due to its enrichment in light ion lithophile elements (LILE), particularly Cs, Rb, Ba and K (Figure 5b).

The eclogitic schist (#OSR-02C) has Al₂O₃, Mg#, CaO and Na₂O + K₂O compositions comparable to those of the ‘common’ eclogites from the Bughea Complex, but shows higher SiO₂ and lower TiO₂ contents (Figure S15.1). Its REE profile shows a pronounced LREE depletion relative to the continental crust and the ocean island basalt (OIB) but enrichment relative to mid-ocean ridge basalts (MORB) and a slight depletion in HREE relative to both the continental crust and MORB. This pattern is similar to that from trench sediments in the Vanuatu Arc (Plank and Langmuir 1998) (Figure 5a). In the mantle-normalized trace element profile, this rock displays Rb, Ba, Th and U patterns

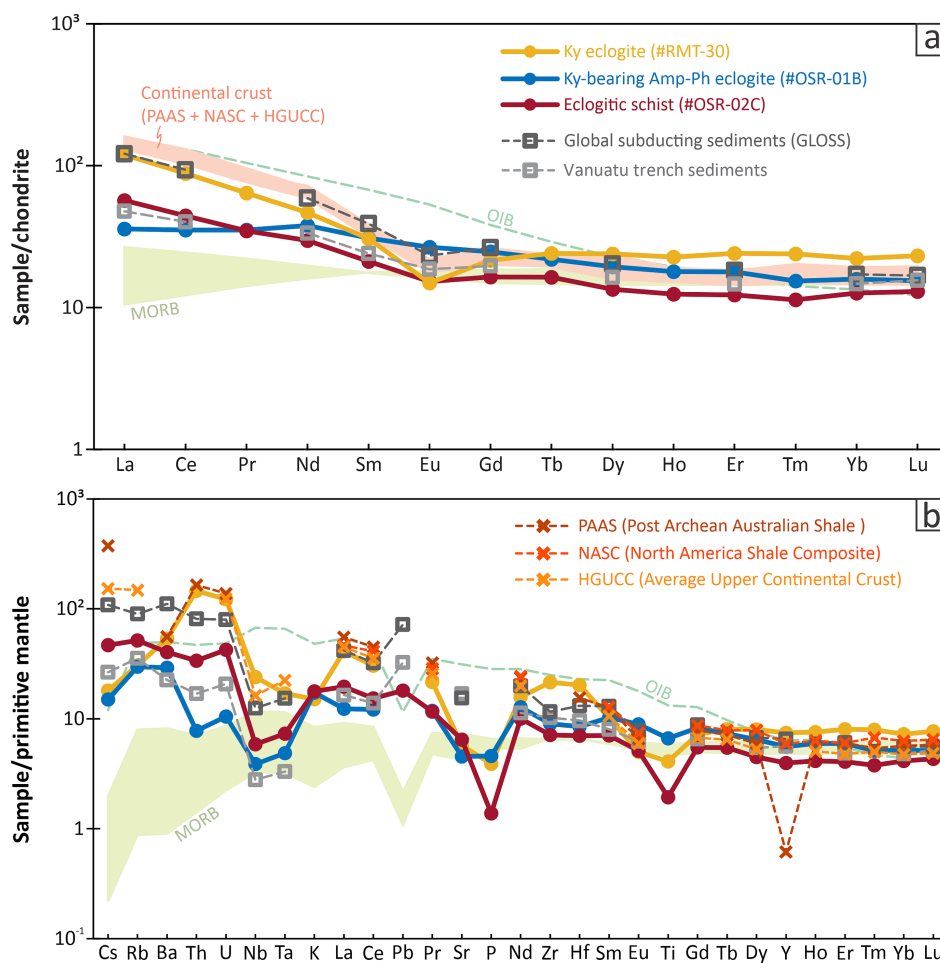


FIGURE 5 | (a) Chondrite-normalized and (b) primitive mantle-normalized diagrams, depicting the trace elements composition of the study eclogites, the continental crust (PAAS from Nance and Taylor 1976, NASC from Taylor and McLennan 1985 and HGUC from Hu and Gao 2008), the oceanic crust (N-MORB, E-MORB and OIB from Sun and McDonough 1989), and global subducting sediments (GLOSS) and Vanuatu trench sediments (Plank and Langmuir 1998). Chondrite and primitive mantle normalizing values are from Sun and McDonough (1989).

comparable to those of OIB and exhibits pronounced negative P and Ti anomalies. The Zr and Hf contents are similar to those of MORB (Figure 5b).

4.3 | Phase Equilibrium Modelling and Zr-in-Rutile Thermometry

4.3.1 | Kyanite Eclogite (#RMT-30)

Four metamorphic stages were determined for the kyanite eclogite (#RMT-30) (Figure 6a). The first phase diagram was calculated using the whole-rock bulk composition, representing the early prograde conditions (stage M_1). The observed paragenesis is stable over a broad P - T range of $\sim 490^\circ\text{C}$ - 740°C and 1.4-2.3 GPa and the overlap of Grt_1 compositional isopleths (X_{alm} , X_{prp} and X_{grs}) constrains equilibration at $\sim 610^\circ\text{C}$ - 650°C and 1.5-1.6 GPa (Figure S8.1).

Late prograde conditions (stage M_2) were calculated using a reactive bulk composition by subtracting the composition of Grt_1 from the whole-rock bulk composition. The observed

paragenesis is stable across a P - T range of $\sim 600^\circ\text{C}$ - 740°C and 1.6-2.3 GPa and the overlap of Grt_2 compositional isopleths (X_{alm} , X_{prp} and X_{grs}) constrains equilibration at $\sim 665^\circ\text{C}$ - 735°C and 1.8-2.1 GPa (Figure S8.2).

The third phase diagram was calculated by subtracting the compositions of garnet core and mantles from the whole-rock bulk composition, reflecting minimum peak conditions (stage M_3). The observed paragenesis is stable over a broad P - T range ($T > 500^\circ\text{C}$; $P > 2.0$ GPa) and the overlap of Grt_4 (X_{alm} , X_{prp} and X_{grs}) and phengite (Si c.p.f.u) compositional isopleths constrains equilibration at $\sim 700^\circ\text{C}$ - 820°C and 3.2-3.6 GPa (Figure 7). For a fixed $P = 3.5$ GPa, the Zr-in-rutile thermometer yields a lower temperature range of $\sim 584^\circ\text{C}$ - 634°C (mean = $604^\circ\text{C} \pm 16^\circ\text{C}$; 1σ , $n = 8$) (Figure 7c).

Retrograde conditions (stage M_4) were constrained using a local bulk composition derived from a single microdomain containing intergrowths of Amp_2 , albite and paragonite partially replacing omphacite-jadeite, kyanite and phengite along their grain boundaries. This paragenesis is stable over a P - T range of $T < 675^\circ\text{C}$; $P < 1.6$ GPa (Figure S8.3).

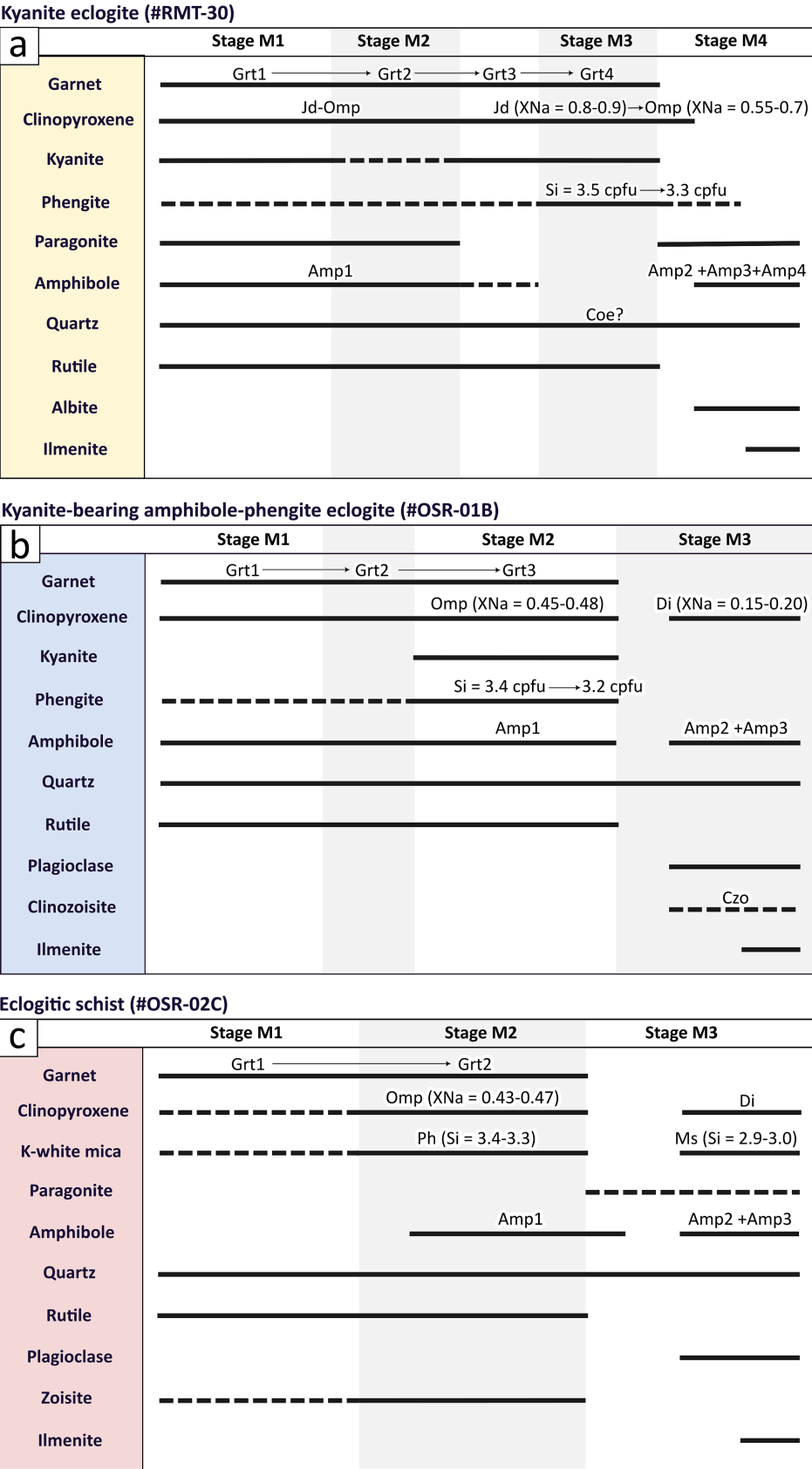


FIGURE 6 | Synthesis of the paragenetic sequence of the (a) kyanite eclogite (#RMT-30), (b) kyanite-bearing amphibole-phengite eclogite and (c) quartz-rich zoisite-phengite eclogite (#OSR-02C) as inferred from textural and mineral chemical features. Continuous lines represent the certain stability of mineral phases, whereas dashed lines indicate phases whose presence during the evolution stage is uncertain.

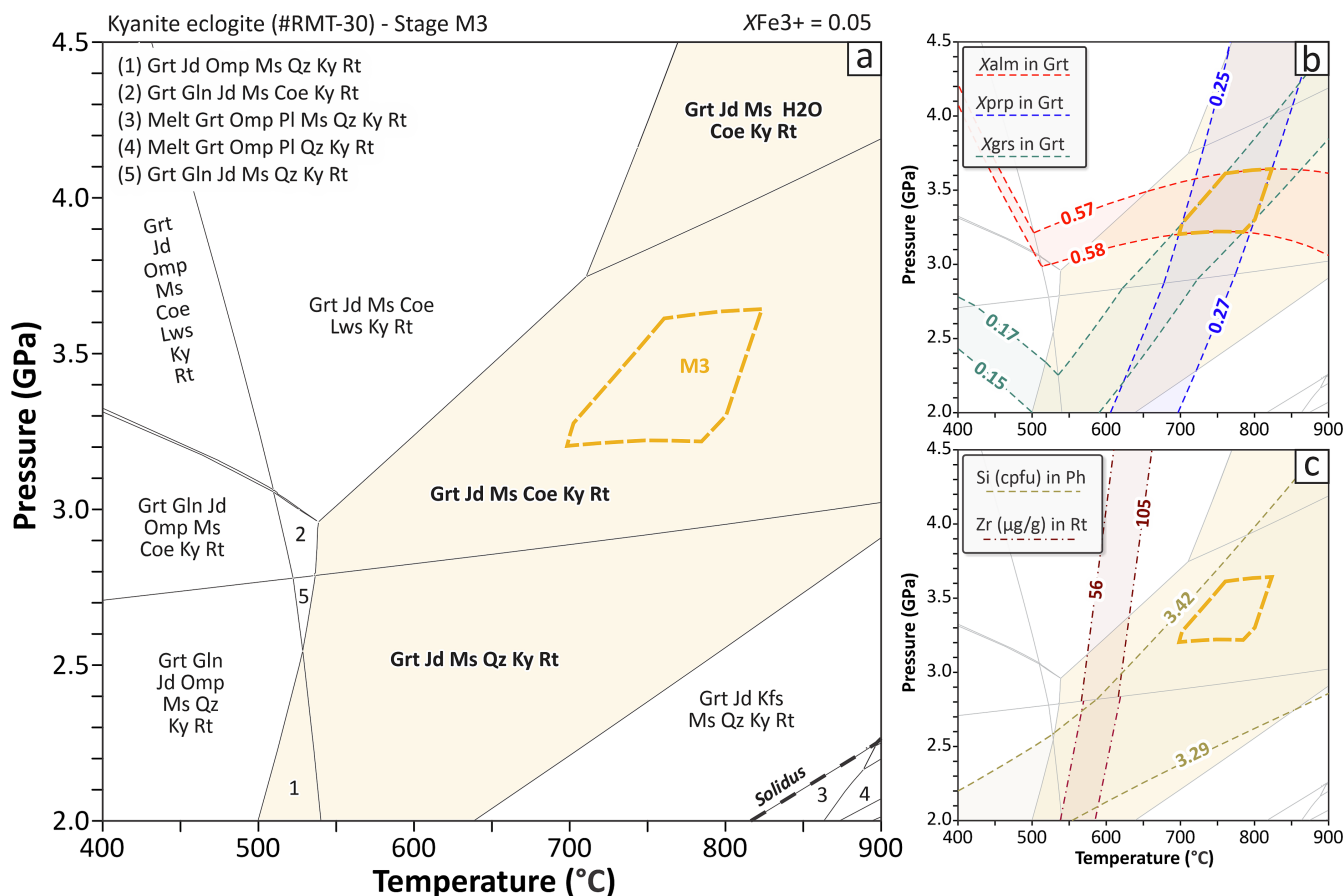


FIGURE 7 | (a) Isochemical P - T phase diagram calculated for stage M_3 of the kyanite eclogite (#RMT-30) using the fractionated bulk composition. The stability fields of the peak paragenesis are highlighted in light yellow. The dashed area (M_3) represents the overlap of compositional isopleths of (b) garnet (X_{alm} , X_{prp} and X_{grs}) and (c) phengite (Si c.p.f.u.), which constraint minimum peak pressure conditions at $\sim 700^\circ\text{C}$ - 820°C and 3.2–3.6 GPa. Compositional isopleths of Zr ($\mu\text{g/g}$) in rutile are also depicted.

4.3.2 | Kyanite-Bearing Amphibole-Phengite Eclogite (#OSR-01B)

Three metamorphic stages were determined for the kyanite-bearing amphibole-phengite eclogite (#OSR-01B) (Figure 6b).

The first phase diagram was calculated using the measured whole-rock bulk composition, reflecting early prograde conditions (stage M_1). The prograde paragenesis could not be robustly determined given the lack of diagnostic phases included in Grt_1 , except for amphibole, omphacite and quartz. The overlap of compositional isopleths of Grt_1 (X_{prp} and X_{grs}) and rutile inclusions (Zr $\mu\text{g/g}$) constraints equilibration at $\sim 590^\circ\text{C}$ - 615°C and 1.7–1.8 GPa, within a stability field where biotite is stable in ~ 3 vol.% (Figure S9.1).

Minimum peak conditions (stage M_2) were calculated by subtracting the composition of garnet core and mantle from the whole-rock bulk composition. The observed paragenesis is stable across a broad P - T range ($T < 700^\circ\text{C}$ and $P > 1.8$ GPa) and the overlap of compositional isopleths of Grt_3 (X_{alm} , X_{prp} and X_{grs}), phengite (Si c.p.f.u.) and rutile (Zr $\mu\text{g/g}$) constraints equilibration at $\sim 580^\circ\text{C}$ - 610°C and 2.0–2.2 GPa (Figure 8).

Late retrograde conditions (stage M_3) were constrained using a phase diagram based on a local bulk composition derived from a single microdomain containing both Amp_2 coronas around garnet and Amp_3 +plagioclase symplectites, which partially replace omphacite and phengite along their grain boundaries. This retrograde assemblage is stable over a restricted P - T range of $\sim 585^\circ\text{C}$ - 615°C and ~ 1.3 GPa, in which titanite is present in modal proportions inferior to 1 vol.% (Figure S9.2).

4.3.3 | Eclogitic Schist (#OSR-02C)

Three metamorphic stages were identified for the eclogitic schist (#OSR-02C). However, due to the absence of diagnostic inclusions, the prograde assemblage remains undetermined and the lack of robust intersections of garnet compositional isopleths prevents a reliable constraint on prograde conditions.

Peak conditions (stage M_2) were calculated by subtracting the composition of garnet cores from the whole-rock bulk composition. The inferred paragenesis is stable over a P - T range of $\sim 520^\circ\text{C}$ - 775°C and 1.2–2.4 GPa and the intersection of Grt_2

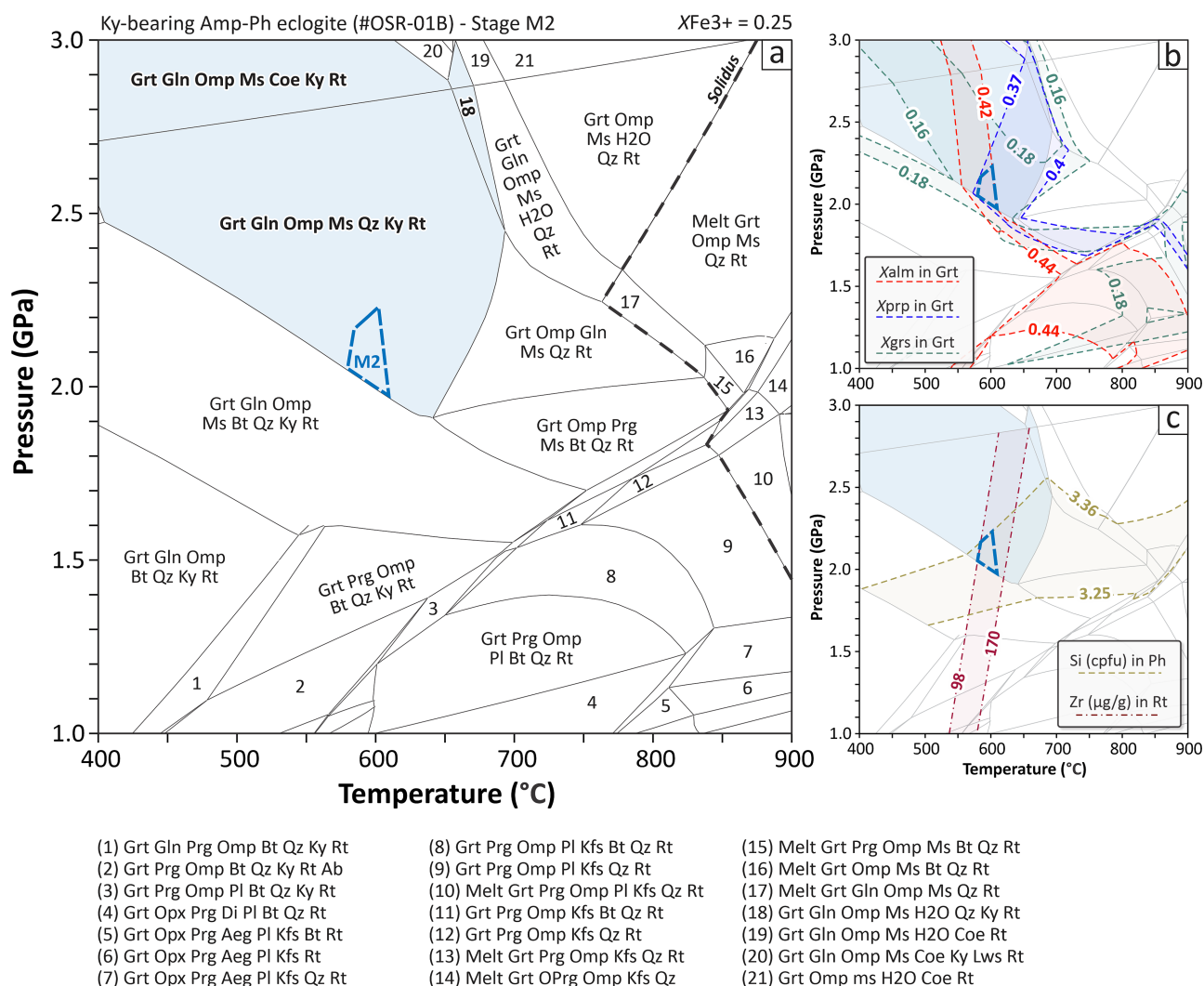


FIGURE 8 | (a) Isochemical P - T phase diagram calculated for stage M_2 of the kyanite-bearing amphibole-phengite eclogite (#OSR-01B) using the fractionated bulk composition. The stability fields of the peak paragenesis are highlighted in light blue. The dashed area represents the overlap of compositional isopleths of (b) garnet (X_{alm} , X_{prp} and X_{grs}), (c) phengite (Si c.p.f.u.) and rutile (Zr $\mu g/g$), which constraint minimum peak pressure conditions at $\sim 580^{\circ}C$ – $610^{\circ}C$ and 2.0–2.3 GPa.

(X_{alm}), phengite (Si c.p.f.u.) and rutile (Zr $\mu g/g$) compositional isopleths constraint equilibration at $\sim 530^{\circ}C$ – $600^{\circ}C$ and 1.9–2.4 GPa (Figure 9). Because P - T conditions could not be robustly constraint from the intersection garnet compositional isopleths, conventional garnet-omphacite thermometry was additionally applied using the calibration of Powell (1985). For these calculations, analyses of close-related pyrope-rich garnet rims (Grt₂) and Na-rich omphacite domains were selected. Assuming a fixed pressure of 2.4 GPa, the garnet-clinopyroxene pairs yield temperatures of $591^{\circ}C \pm 22^{\circ}C$ to $607^{\circ}C \pm 26^{\circ}C$, consistent with the temperature range constrained by the overlap of compositional isopleths in the calculated phase diagram.

Retrograde conditions (stage M_3) were calculated using a local bulk composition derived from a single microdomain containing both Amp₂ + albite-oligoclase symplectites, which partially replace omphacite and phengite, as well as Amp₃ coronas around garnet. This retrograde assemblage is stable over a P - T range of $\sim 600^{\circ}C$ – $615^{\circ}C$ and 1.0–1.2 GPa (Figure S9.1).

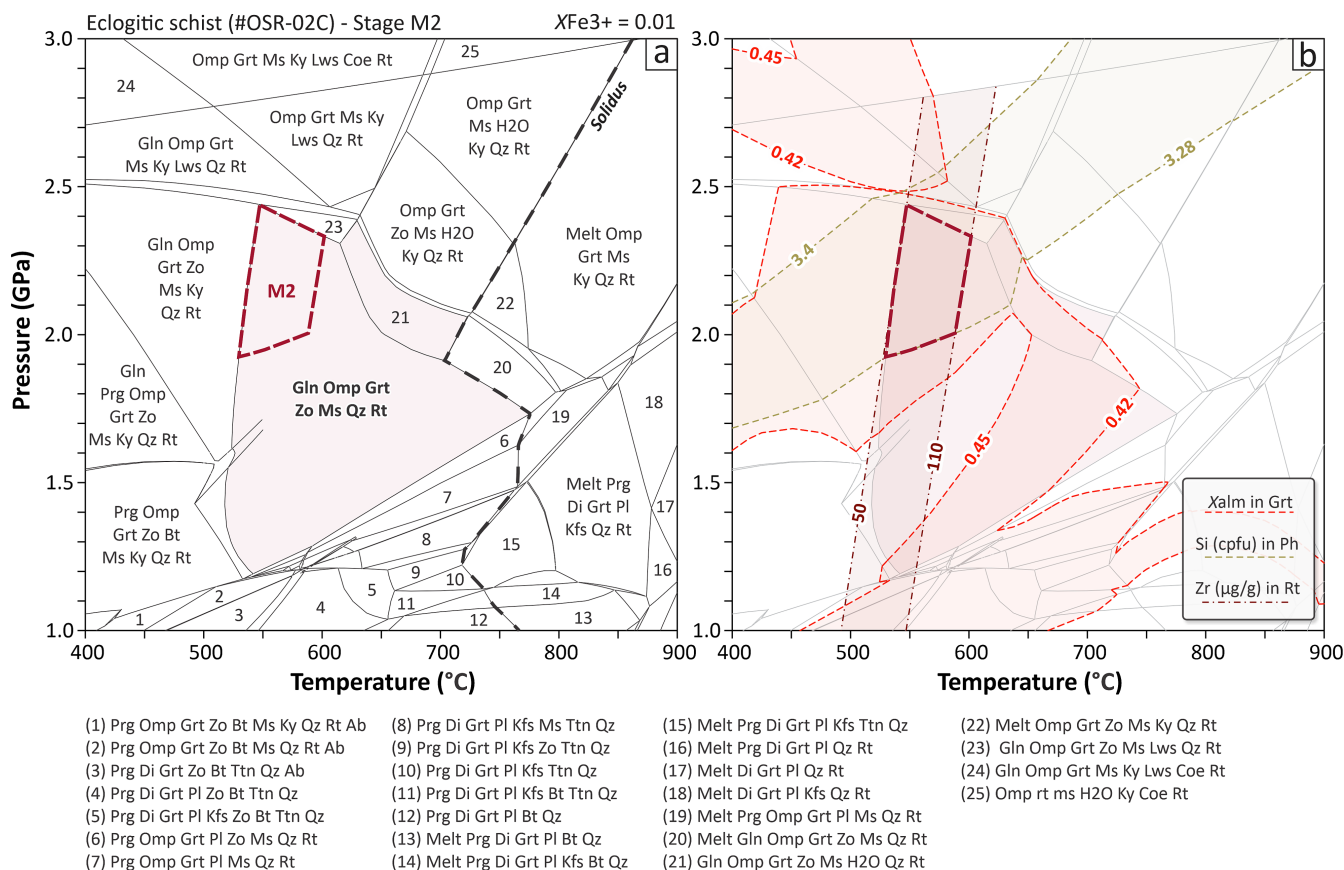
4.4 | Trace Element and Geochronological Data

4.4.1 | Zircon

A total of 117 zircon grains from the kyanite eclogite (#RMT-30) were analysed for U-Pb geochronology and trace-element chemistry. Whenever possible, both cores and rims of the same grain were analysed; however, owing to their fine grain size, only cores could be analysed in most cases.

The zircon grains display heterogeneous characteristics in both morphology and cathodoluminescence (CL) response (Figure 10a,b). Grain shapes range from equidimensional to prismatic, with pyramidal to rounded terminations. They are dominated by cores that either show moderate luminescence with oscillatory zoning or bright luminescence with a more uniform appearance. Thin rims with variable CL responses are locally observed but were too small to be analysed by LA-ICP-MS.

Zircon cores ($n=66$) yield scattered concordant ages, with most data yielding Neoproterozoic to Cambrian ages (700–500 Ma;



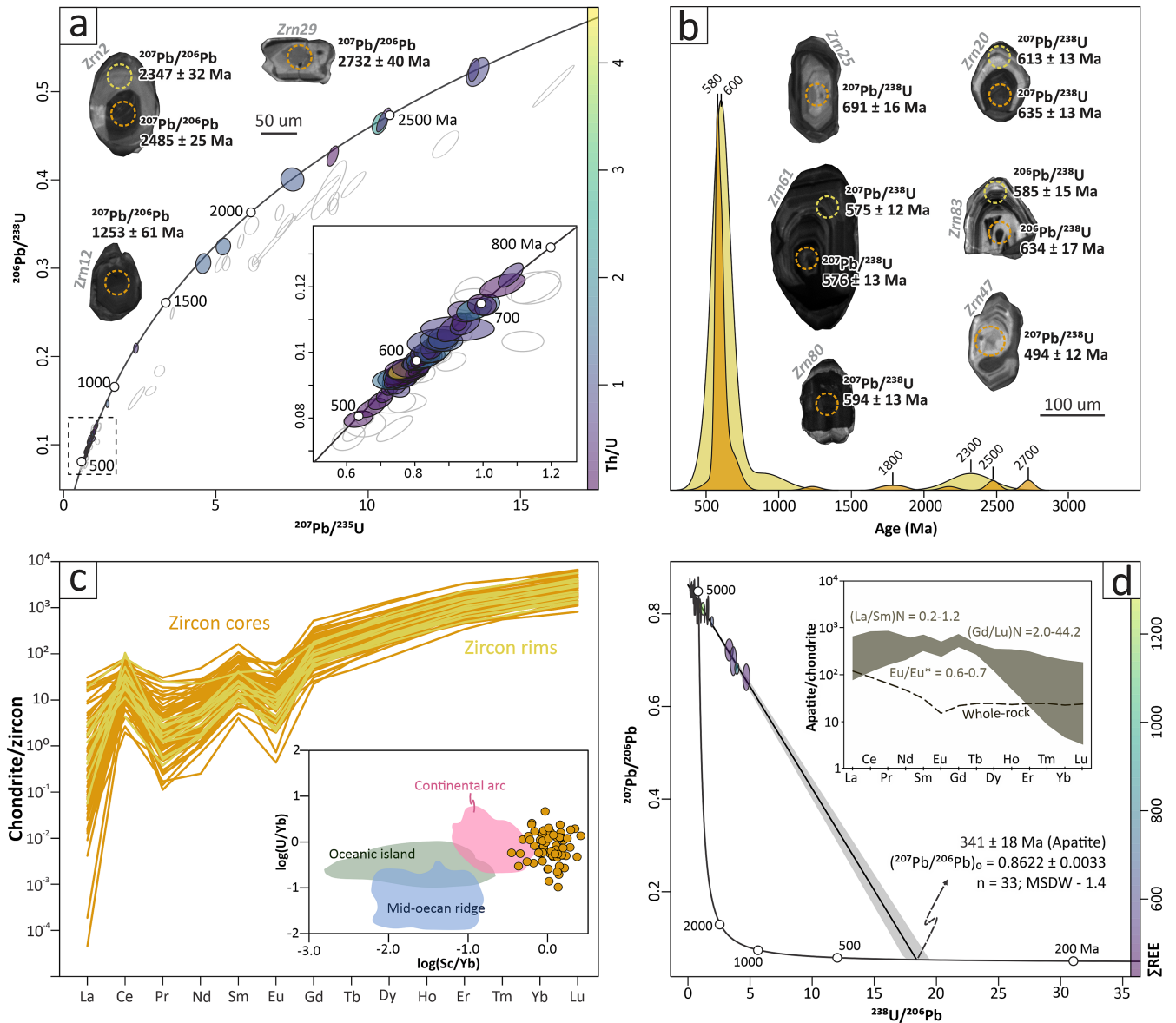


FIGURE 10 | Summarized zircon and apatite U–Pb geochronological data from the kyanite eclogite (#RMT-30). **(a)** Concordia diagram depicting concordant (filled coloured ellipses) and discordant (hollow grey ellipses) zircon U–Pb data. **(b)** Kernel density estimation (KDE) diagram depicting the distribution of zircon U–Pb ages and cathodoluminescence images of representative grains. Ages from the core are depicted in orange, whereas ages from the rims are depicted in dark yellow. **(c)** REE profile of all concordant zircon analyses in a chondrite–normalized diagram. The inset shows a discriminant $\log(\text{U}/\text{Yb})$ vs. $\log(\text{Sc}/\text{Yb})$ diagram from Grimes et al. (2015) depicting the continental arc compositions of all concordant zircon cores. **(d)** Inverse U–Pb isochron in the Tera–Wasserburg diagram produced from in situ isotopic analyses of apatite in the kyanite eclogite. The inset shows a chondrite–normalized apatite and whole-rock REE profile. Chondrite normalizing values are from Sun and McDonough (1989). All ellipses of the figure represent 2σ uncertainties.

5 | Discussion

5.1 | Metamorphic Evolution of Eclogites From the Bughea Complex

5.1.1 | Kyanite Eclogite (#RMT-30)

The kyanite eclogite displays normalized REE and trace element profiles that closely resemble those of the continental crust (Figure 5). It records the most complex clockwise metamorphic evolution among the studied eclogites (Figure 12).

The early prograde stages are characterized by increasing temperature and pressure from $\sim 610^\circ\text{C}$ – 650°C and 1.5–1.6 GPa to $\sim 665^\circ\text{C}$ – 735°C and 1.8–2.1 GPa (Figures 12, S8.1 and S8.2). During the prograde path, partial growth of jadeite-omphacite, kyanite and garnet mantles was likely a result of the breakdown of Amp_1 and paragonite through reactions $\text{Pg} = \text{Ky} + \text{Jd} + \text{H}_2\text{O}$ and $\text{Gln} + \text{Pg} = \text{Grt} + \text{Jd} + \text{Qz} + \text{H}_2\text{O}$.

Minimum peak pressure conditions are constrained at $\sim 700^\circ\text{C}$ – 820°C and 3.2–3.6 GPa, indicating substantial pressure and temperature increase to ultrahigh-pressure conditions

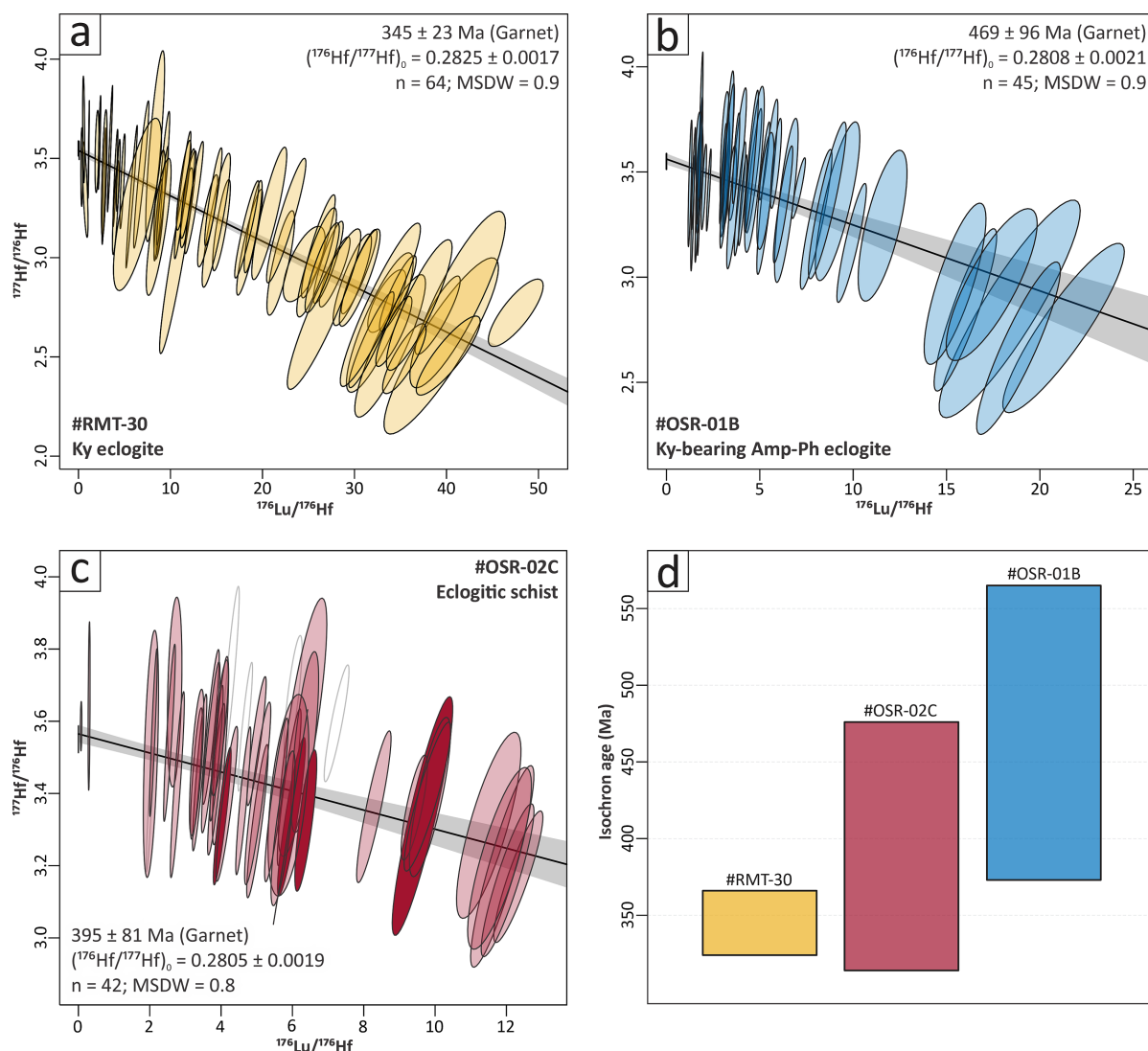


FIGURE 11 | Inverse Lu–Hf isochron produced from in situ isotopic analyses of garnet in the (a) kyanite eclogite (#RMT-30), (b) kyanite-bearing amphibole–phengite eclogite (#OSR-01B) and (c) eclogitic schist (#OSR-02C). (d) Weighted mean ages for garnet growth in all three samples, depicting differences between garnet Lu–Hf ages.

during the final episode of garnet growth (Figures 7 and 12). These conditions correspond to T/P ratios of $\sim 219^\circ\text{C}$ – $245^\circ\text{C}/\text{GPa}$, comparable to modern cold subduction regimes, typically ranging between 150°C and $375^\circ\text{C}/\text{GPa}$ (Brown and Johnson 2019).

Rutile yields lower temperatures than those calculated for prograde and peak pressure conditions based on the intersection of garnet and phengite chemical isopleth (Figures 7c, S8.1 and S8.2). The Zr-in-rutile geothermometer requires equilibrium among rutile, zircon and quartz (Zack et al. 2004). Although these phases are present in the kyanite eclogite, zircon ages are significantly older than the inferred metamorphism, indicating little to no metamorphic zircon growth during the eclogitic metamorphism (see discussion below). This suggests that zircon and rutile may not have attained equilibrium during eclogitic metamorphism, thereby preventing effective Zr buffering. Consequently, rutile temperatures likely underestimate prograde and peak pressure conditions. A similar discrepancy was reported by Kendrick and Indares (2018) in high-pressure granulites from the Grenville Province, Canada.

While having reached P – T conditions to form coesite, no inclusions of it have been observed in the kyanite eclogite (#RMT-30). This may be the result of intense fracturing of the host phases, particularly garnet porphyroblasts, likely releasing the confining pressure and facilitating infiltration of late-stage fluids that induced extensive retrograde re-equilibration (Mosenfelder et al. 2005). The observed radial microcracks surrounding scattered quartz inclusions in garnet, clinopyroxene and kyanite could reflect the former transformation of coesite to quartz during retrogression, as suggested by Săbău (2000). However, this feature is inconclusive, because it could also be the result of decompression from less extreme depths (Whitney et al. 2000).

The retrograde stage is constrained at P – T conditions below $\sim 675^\circ\text{C}$ and 1.6 GPa from intergrowths of Ca-amphibole, albite and paragonite replacing jadeite-omphacite, phengite and kyanite (Figure S8.3). This indicates a clockwise P – T path marked by minor cooling and significant decompression, which could be a result of exhumation to depths below $\sim 53 \text{ km}$ (assuming an average lithostatic gradient of $\sim 33.33 \text{ km/GPa}$) driven

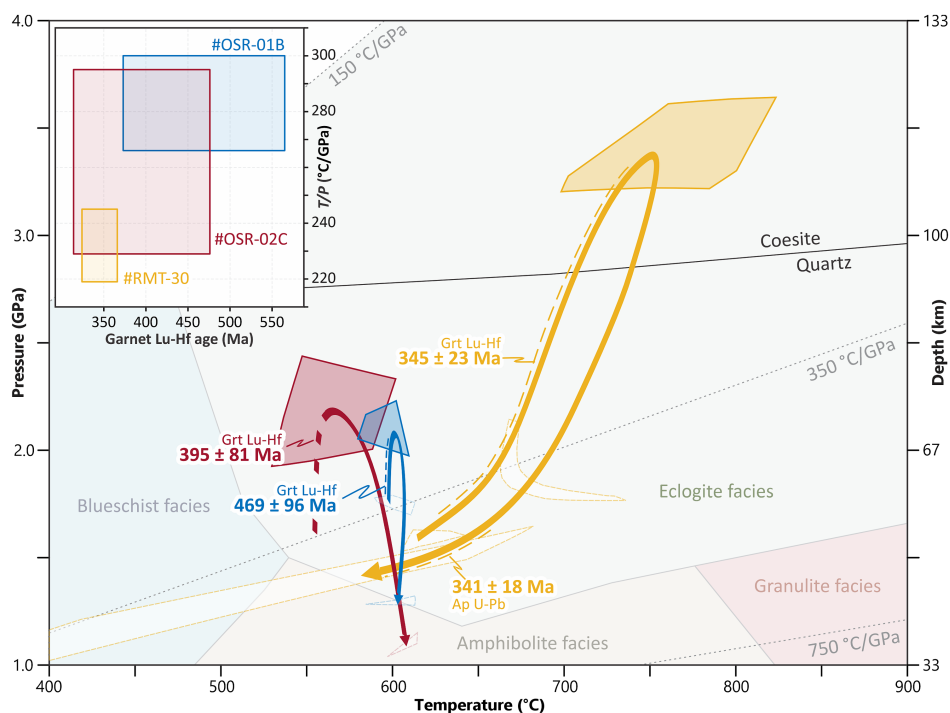


FIGURE 12 | Diagram summarizing the reconstructed P - T - t paths for all three eclogitic rocks from the Bughea Complex. Metamorphic facies fields are redrawn after Wei and Duan (2019). The pressure to depth conversion was carried out assuming an average lithostatic gradient of ~ 33.33 km/GPa. The inset depicts variation of the thermobaric ratios (temperature/pressure, T/P) for the study samples plotted against their garnet Lu-Hf ages.

by circulatory flow within the subduction zone (Platt 1993; Warren 2013; Figure 12).

The P - T path reconstructed for the kyanite eclogite is comparable to those reported for other ultrahigh-pressure metasedimentary rocks from different orogens, including: (i) diamond- and coesite-bearing paragneiss from the Dabie-Sulu belt (~ 3.5 – 4.0 GPa, 750°C – 850°C), showing postpeak near-isothermal decompression followed by cooling and yielding metamorphic ages of 235–225 Ma (Liu and Liou 2011); (ii) coesite-bearing schists from the Western Alps (~ 2.6 – 2.9 GPa, 500°C – 630°C), which record postpeak near-isothermal decompression and metamorphic ages of 60–45 Ma (Reinecke 1998; Manzotti et al. 2025); and (iii) diamond-bearing schist from the Rhodope Complex (~ 3.0 – 3.5 GPa, 700°C – 750°C), which also record near-isothermal decompression followed by cooling (Schmidt et al. 2010), presenting metamorphic ages either around 200 Ma, inferred from associated diamond-bearing paragneiss (Petrik et al. 2016), or around 45–40 Ma, inferred from associated eclogites (Kirchenbaur et al. 2012).

5.2 | Other Eclogitic Rocks

The kyanite-bearing amphibole-phengite eclogite (#OSR-01B) shows basic composition, consistent with other ‘common’ eclogites in the Bughea Complex (Figure S15.1). Its normalized REE and trace element profiles (Figure 5), together with a low Th/La ratio (0.8), are consistent with altered MORBs compositions. The eclogitic schist (#OSR-02C) exhibits a normalized REE and trace-element signatures comparable to sedimentary protoliths

containing volcanic detrital input, similar to trench sediments from the Vanuatu subduction zone (Figure 5).

Petrographic evidence, mineral compositions and thermobarometric estimates indicate that both eclogites record clockwise P - T paths (Figure 12). The calculated prograde and peak pressure conditions for the basic eclogite indicate near-isothermal burial, with pressure increasing from 1.7–1.8 to 2.0–2.3 GPa at a similar temperature range of $\sim 580^{\circ}\text{C}$ – 615°C (T/P ratios of $\sim 266^{\circ}\text{C}$ – $300^{\circ}\text{C}/\text{GPa}$; Figures 8 and S9.1). The eclogitic schist records broadly similar peak pressure conditions, around 530°C – 600°C at 1.9–2.3 GPa (T/P ratios of $\sim 250^{\circ}\text{C}$ – $280^{\circ}\text{C}/\text{GPa}$; Figure 9), consistent with cold subduction regimes. The reconstructed retrograde paths indicate near-isothermal decompression to $\sim 585^{\circ}\text{C}$ – 615°C and 1.3 GPa for the basic eclogite, whereas the eclogitic schist records minimal heating during decompression to $\sim 600^{\circ}\text{C}$ – 615°C and 1.0–1.2 GPa, corresponding to exhumation to depths of ~ 43 – 33 km (Figures S9.2, S10.1 and 12).

5.3 | Age Significance

5.3.1 | Zircon Detrital Signature

Zircon cores from the kyanite eclogite (#RMT-30) yield scattered concordant ages spanning from the Archean to the Cambrian, with a dominant population between 700 and 500 Ma, whereas the rims display a similar age distribution, with a main population between 650 and 500 Ma (Figure 10a,b). These ages are unlikely related to the ultrahigh-pressure metamorphism, as most REE patterns differ significantly from those reported for typical

eclogitic zircon (Rubatto 2017 and references therein). These differences include the following: (i) high total REE contents, particularly for HREE, contrasting with the more depleted patterns expected for eclogitic zircon formed in equilibrium with other HREE-bearing phases, such as garnet; (ii) pronounced negative Eu anomalies in chondrite-normalized diagrams, which are uncommon in eclogitic zircon due to the general absence of plagioclase during high-pressure metamorphism; and (iii) elevated Th/U ratios, which are usually low (<0.1) for metamorphic zircon (Figure 10c). These features, coupled with elevated U/Yb and Sc/Yb ratios, indicate that most of the zircon in the kyanite eclogite is likely igneous, formed in a continental-arc setting (Grimes et al. 2015; Kirkland et al. 2015; Rubatto 2017).

The spreading of the zircon ages and their chemical signatures suggest either an igneous protolith for the kyanite eclogite with a long-lived crystallization and hosting inherited Proterozoic and Neoproterozoic xenocrysts or a sedimentary protolith dominated by Neoproterozoic-Cambrian sources, with a maximum depositional age of ca. 500 Ma (Figure 10a,b). Considering the heterogeneous textural and chemical characteristics of the zircon grains, the broad spread of Neoproterozoic-Cambrian ages of 200–150 Myr, which is unlikely for the crystallization of a single igneous body (e.g., Coleman et al. 2004), and the geochemical signature of the rock, a sedimentary precursor is favoured and consistent with previous interpretations (cf., Negulescu et al. 2009).

Accordingly, the range of ages, with dominant populations between 700 and 500 Ma, the maximum depositional age of ca. 500 Ma, the zircon geochemical signature typical of a continental-arc setting, and the small age difference between the youngest core (ca. 495 Ma) and the youngest rim (ca. 515 Ma) together suggest that the sedimentary protolith was deposited in an arc-related trench or forearc basin (Cawood et al. 2012). This scenario is plausible since similar ages have been reported for Cadomian to early Variscan arc magmatism across the basement of the South Carpathians (Balintoni et al. 2014 and references therein).

5.3.2 | Timing of Eclogitic Metamorphism

The kyanite eclogite (#RMT-30) yields a relatively well-constrained garnet Lu–Hf age of 345 ± 23 Ma (Figure 11a), interpreted as the timing of subduction-related prograde metamorphism. This age is consistent with the apatite U–Pb age of 341 ± 18 Ma (Figure 10d). The low total REE contents in apatite indicate metamorphic growth and/or fluid-induced depletion (O'Sullivan et al. 2020; Ribeiro et al. 2020). Given the significant influence of hydrous fluids during retrogression, evidenced by the widespread occurrence of retrograde hydrous assemblages replacing earlier anhydrous phases, and considering that the closure temperature of U–Pb geochronometer in apatite is significantly lower ($\sim 350^\circ\text{C}$ – 600°C) than that of Lu–Hf in garnet (up to $\sim 800^\circ\text{C}$), the apatite age is interpreted as the timing of cooling. Furthermore, the discrete negative Eu anomaly in apatite may reflect the bulk-rock composition or, alternatively, re-equilibration with plagioclase during the retrograde path (Henrichs et al. 2018). This latter scenario is favoured by the elevated HREE contents relative to the bulk rock, which may reflect HREE release during partial garnet dissolution and subsequent

incorporation into apatite during retrogression, consistent with pressures below ~ 1.6 GPa.

Similar muscovite and biotite Ar–Ar ages of 360–335 Ma have been reported for mylonitic gneisses along major tectonic boundaries of the Bughea Complex, as well as phengite and amphibole Ar–Ar ages of 355–340 Ma for a retrogressed eclogite (Axente et al. 2008), reflecting the timing of development of mylonitic fabrics and retrograde assemblages during exhumation and cooling. Accordingly, we interpret the apatite ages as reflecting the timing of exhumation of the kyanite eclogite. The overlap between garnet and apatite ages likely indicates relatively rapid subduction and exhumation, consistent with previous interpretations (Axente et al. 2008).

The metamorphic ages obtained for the kyanite eclogite are broadly consistent with those reported for other eclogites in the central part of the South Carpathians (360–340 Ma; Medaris Jr. et al. 2003) and across other Variscan terrains, including the Alps (335–330 Ma; Giacomini et al. 2007; Herwartz et al. 2011), the Rhodope Metamorphic Complex (ca. 335 Ma; Trapp et al. 2021), the Armorican Massif (350–345 Ma; Tual et al. 2025), the Bohemian Massif (370–335 Ma; Schmädicke et al. 1995; Bröcker et al. 2009) and the French Massif Central (380–350 Ma; Lotout et al. 2018; de Hoÿm de Marien et al. 2023).

The basic eclogite (#OSR-01B) yielded an older metamorphic age (469 ± 96 Ma) than the kyanite eclogite, whereas the eclogitic schist (#OSR-02C) yielded an age of 395 ± 81 Ma, which is comparable to the other samples within uncertainty (Figure 11). This indicates that garnet growth was not coeval along the Bughea Complex, which could be a result of two possible scenarios.

The first scenario is a prolonged eclogitic metamorphism in a relatively long-lived subduction during the early to late stages of the Variscan Orogeny. In this scenario, the basic eclogite would record an earlier stage of subduction, whereas the other eclogites were subducted later during continued convergence. Such a scenario is plausible, as metamorphic ages older than 380 Ma have been reported for Variscan-related eclogites elsewhere (Stosch and Lugmair 1990; Giacomini et al. 2007; Lotout et al. 2018), and similar evidence of such prolonged subduction has been documented in the Franciscan Complex (Mulcahy et al. 2018) and in the French Massif Central (de Hoÿm de Marien et al. 2023 and references therein).

The second scenario is multiple episodes of eclogite facies metamorphism produced by distinct events. Comparable zircon U–Pb ages of 590–550 and 480–450 Ma have been reported for Cadomian and pre-Variscan arc magmatism across the basement of the South Carpathians (Balintoni et al. 2014 and references therein) and could be associated with the formation of older, pre-Variscan eclogites.

Given that all eclogites in the easternmost South Carpathians occur within a relatively narrow and coherent unit, and considering the lack of other robust evidence of pre-Variscan regional eclogite facies metamorphism in the South Carpathians, the interpretation of a single prolonged Variscan subduction is favoured. Nevertheless, more robust geochronological constraints are required to support further interpretations.

5.4 | Tectonic Significance

The Variscan Orogeny resulted from the diachronous amalgamation of the northern margin of Gondwana, the Peri-Gondwanan microcontinental terranes and the southern margin of Laurussia over ca. 200 Myr during the Cambrian–Carboniferous (Franke et al. 2021; Schulmann et al. 2025 and references therein). In the South Carpathians, Variscan high-pressure metamorphism is mostly recorded within the Armorica-type Peri-Gondwanan terrane of the Getic Unit (Iancu and Seghedi 2017 and references therein).

Based on metamorphic and geochronological similarities between the eclogites from the Bughea Complex and Variscan eclogites elsewhere (e.g., Trapp et al. 2021), we interpret these rocks to have formed in a comparable tectonic setting, likely related to the closure of the Rheic Ocean during Devonian–Carboniferous convergence and amalgamation of Peri-Gondwanan terranes. In this scenario, the protolith of the basic eclogite may represent a fragment of altered Rheic oceanic crust, whereas the eclogitic schist likely derives from sediments deposited in volcanic arc-related basin and subsequently accreted during ongoing convergence. The kyanite eclogite is interpreted to derive from sediments deposited along the active margin of a Peri-Gondwanan terrane, possibly within an accretionary prism or trench setting.

5.5 | Conditions for the Recording of Different Metamorphic Evolutions in the Bughea Complex

The continental-derived kyanite eclogite (#RMT-30) records higher peak pressure conditions than the oceanic- and volcanic arc-derived rocks (#OSR-01B and #OSR-02C, respectively). Two scenarios may explain this contrast: (i) all three samples followed broadly similar metamorphic evolutions, but only the kyanite eclogite preserved the complete P – T record, or (ii) the eclogites underwent distinct trajectories through the lithosphere, with the kyanite eclogite reaching greater depths.

The first scenario implies that the kyanite eclogite preserves a more complete record of the metamorphic evolution, whereas the other eclogites retain only partial records. This could result from differences in reaction kinetics and/or retrograde resetting. Reaction kinetics are usually enhanced in metasedimentary rocks, largely due to bulk composition and fluid availability, which favour the development of chemically zoned porphyroblasts and diverse parageneses capable of recording a more complete metamorphic evolution (Rubie 1986; Carlson 2011). In addition, garnet in typical basic eclogites is less likely to record peak pressure conditions due to fractional crystallization and bulk-rock depletion, which can hamper garnet growth during late prograde to peak stages (Konrad-Schmolke et al. 2008). However, the role of reaction kinetics in controlling variations in the P – T record is difficult to robustly assess and remains uncertain. Alternatively, contrasting retrograde resetting appears unlikely as it would require substantial fluid influx and/or heating during retrogression (Heinrich 1982; Proyer 2003; Walsh and Hacker 2004; Fitzherbert et al. 2005; Young and Kylander-Clark 2015), neither of which are supported by textural evidence or by the reconstructed P – T paths.

The second scenario is plausible considering the *mélange* setting proposed for the Bughea Complex based on lithological association, petrological investigations and geochemical data (Iancu et al. 1998; Săbău 2000; Negulescu and Săbău 2015). In subduction-related *mélanges*, eclogitic blocks can be transported from different depths across the subduction interface, particularly in the presence of enclosing rocks with low effective viscosity, promoting ductile flow, long-distance transport and juxtaposition of rigid blocks that record distinct P – T evolutions via, e.g., underplating (or tectonic erosion) and/or *mélange* flow (Festa et al. 2010; Harvey et al. 2021 and references therein). This is consistent with the lithological association described in the Bughea Complex, where distinct eclogitic blocks are hosted by sheet silicate-rich rocks (i.e., dominantly semipelitic micaschist and mica gneiss), which are likely derived from an accretionary wedge setting (Săbău 2000; Negulescu and Săbău 2015). Similar lithological associations have been reported in other subduction-related *mélanges* elsewhere, including in the Franciscan Complex of California (metamorphic age of 175–160 Ma; Mulcahy et al. 2018), in the South Tien Shan suture zone of Kyrgyzstan (320–315 Ma; Loury et al. 2018) and in the Central Asian Orogenic Belt (ca. 410 Ma; Wang et al. 2017).

In this context, the distinct metamorphic conditions recorded by the studied samples may reflect variations in subduction geodynamics, as changes in thermal regimes and convergence rates through space and time can influence burial depths and timescales (Wortel 1984; Agard et al. 2009, 2016; Soret et al. 2022; Turino and Holt 2024). The reconstructed P – T paths depict decreasing peak T/P ratios over time (from 266–300 to 219°C–245°C/GPa; Figure 12), suggesting a transition from relatively warmer regimes associated with slower, shallower subduction to colder conditions favouring deeper and possibly faster subduction, similar to the scenario described by Agard et al. (2018) and Soret et al. (2022). In addition, the buoyant nature of continental crust likely facilitated exhumation of the kyanite eclogite from greater depths relative to denser basic compositions (Ernst et al. 1997; Hermann 2002; Warren 2013; Chapman et al. 2019).

6 | Conclusions

This study shows that lithologically diverse eclogites within a modern-style Variscan *mélange* (i.e., the Bughea Complex) in the easternmost South Carpathians record diachronous cold subduction with distinct P – T evolutions.

All three eclogites underwent clockwise P – T trajectories. The kyanite eclogite (#RMT-30) has a composition compatible with the continental crust and records the highest peak pressure conditions, consistent with ultrahigh-pressure metamorphism (~3.2–3.6 GPa at 700°C–820°C). The other eclogites derive from altered MORB (#OSR-01B) and trench sediments in volcanic arc settings (#OSR-02C) and record lower peak pressure conditions of ~1.9–2.4 GPa at 530°C–610°C. The retrograde evolution of all eclogites is characterized by decompression to pressures below ~1.7 GPa over moderate temperatures (~600°C), consistent with relatively rapid exhumation typical of modern-day cold subduction systems, possibly driven by circulatory flow within the subduction system.

The heterogeneous ages and trace-element signatures of zircon from the kyanite eclogite indicate a detrital origin derived from continental-arc and/or cratonic sources, with deposition in a trench or forearc basin and maximum depositional age of ca. 500 Ma. Garnet Lu-Hf ages indicate diachronous metamorphism in the Bughea Complex during Variscan subduction, whereas the apatite U-Pb age of ca. 340 Ma likely reflects cooling and exhumation. These ages are consistent with those reported in other eclogites across the Variscan Orogen and suggest a relatively prolonged subduction during the closure of the Rheic Ocean.

The distinct P - T - t evolutions recorded by these eclogites likely reflect the rheology of the mélange, variations in the subduction geodynamics and composition-related buoyancy, which seems to favour exhumation of continental-derived rocks. Limited thermal overprint during retrogression also seems critical for preserving peak pressure conditions, particularly in nonbasic rocks.

This study highlights that continental metasedimentary rocks can be particularly effective at providing information on their sedimentation setting and on recording extreme pressure conditions. Continued investigation of diverse high-pressure rocks, in addition to basic eclogites, is therefore critical for a comprehensive understanding of complex subduction and orogenic systems.

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Conflicts of Interest

The authors declare no conflicts of interest.

Data Availability Statement

The data that support the findings of this study are available in the [Supporting Information](#).

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Supporting Information

Additional supporting information can be found online in the Supporting Information section. **Table S1:** Whole-rock geochemical analyses from the studied rocks and others from the Bughea Complex. **Table S2:1** EPMA amphibole chemical data. **Table S2:2** EPMA biotite chemical data. **Table S2:3** EPMA clinopyroxene chemical data. **Table S2:4** EPMA zoisite/clinozoisite chemical data. **Table S2:5** EPMA garnet chemical data. **Table S2:6** EPMA feldspar chemical data. **Table S2:7** EPMA white mica chemical data. **Table S2:8** EPMA operating conditions. **Table S3:1** EPMA amphibole chemical data. **Table S3:2** EPMA clinopyroxene chemical data. **Table S3:3** EPMA garnet chemical data. **Table S3:4** EPMA feldspar chemical data. **Table S3:5** EPMA white mica chemical data. **Table S3:6** EPMA operating conditions. **Table S4:1** EPMA amphibole chemical data. **Table S4:2** EPMA clinopyroxene chemical data. **Table S4:3** EPMA zoisite chemical data. **Table S4:4** EPMA garnet chemical data. **Table S4:5** EPMA feldspar chemical data. **Table S4:6** EPMA white mica chemical data. **Table S4:7** EPMA operating conditions. **Figure S5:1** Chemical diagrams depicting the composition and classification of (a,b) sodium-calcium and sodium clinopyroxenes according to Morimoto (1988), (c) garnet, (d) plagioclase and (e) phengite in all three eclogites. **Figure S5:2** Chemical diagrams depicting the composition and classification of (a) sodium, (c) sodium-calcium and (d,e) calcium amphiboles in all three samples, according to the classification schemes of Leake et al. (1997) and Hawthorne et al. (2012). **Figure S6:1** Compositional maps of the kyanite eclogite (#RMT-30). (a–e) Map I. (a) Map of mineral phases. Dashed curves represent the approximate boundaries separating garnet chemical zones based on the quantitative compositional maps of (b–e) X_{alm} , X_{prp} , X_{grs} and X_{sps} in the garnet porphyroblast, respectively. (f–j) Map II. (f) Map of mineral phases. Other compositional maps depict chemical variations of (g,h) X_{Na} and X_{Mg} in clinopyroxene and (i) Si (c.p.f.u.) in phengite. **Figure S6:2** Compositional maps of the kyanite-bearing amphibole-phengite eclogite (#OSR-01B). (a) Map of mineral phases. Dashed curves represent the approximate boundaries

separating garnet chemical zones based on the quantitative compositional maps of **(b–e)** X_{alm} , X_{prp} , X_{grs} and X_{sps} in the garnet, respectively. Other compositional maps depict chemical variations of **(f–g)** X_{Na} and X_{Mg} in clinopyroxene, **(h)** Si (c.p.f.u.) in phengite and **(i)** X_{Ca} ($=\text{Ca}^{\text{M4}}/(\text{Ca} + \text{Na})^{\text{M4}}$) in amphibole. **Figure S6:3**—Compositional maps from the eclogitic schist (#OSR-02C). **(a)** Map of mineral phases. Dashed curves represent the approximate boundaries separating garnet chemical zones based on the quantitative compositional maps of **(b–e)** X_{alm} , X_{prp} , X_{grs} and X_{sps} in the garnet, respectively. Other compositional maps depict chemical variations of **(f–g)** X_{Na} and X_{Mg} in clinopyroxene and **(h)** Si (c.p.f.u.) in phengite. **Table S7:1** Rutile chemical data and thermometry (#RMT-30). **Table S7:2** Rutile chemical data and thermometry (#OSR-01B). **Table S7:3** Rutile chemical data and thermometry (#OSR-02c). **Table S7:4** EPMA operating conditions. **Figure S8:1 (a)** Isochemical P – T phase diagrams calculated for stage M_1 of the kyanite eclogite (#RMT-30) using the whole-rock bulk composition. The light-yellow area represents the stability field of the prograde paragenesis. The dashed area represents the overlap of chemical isopleths of **(b)** garnet (X_{alm} , X_{prp} and X_{grs}), which constraint equilibration at $\sim 610^\circ\text{C}$ – 650°C and 1.5–1.6 GPa. Chemical isopleths of Zr ($\mu\text{g/g}$) in rutile included in garnet are also depicted. **Figure S8:2 (a)** Isochemical P – T phase diagrams calculated for stage M_2 of the kyanite eclogite (#RMT-30) using the fractionated bulk composition. The light-yellow area represents the stability field of the prograde paragenesis. The dashed area represents the overlap of chemical isopleths of **(b)** garnet (X_{alm} , X_{prp} and X_{grs}), which constraint equilibration at $\sim 665^\circ\text{C}$ – 735°C and 1.8–2.1 GPa. Chemical isopleths of Zr ($\mu\text{g/g}$) in rutile included in garnet are also depicted. **Figure S8:3 (a)** Isochemical P – T phase diagram calculated for the retrograde stage M_4 of the kyanite eclogite (#RMT-30), using the local bulk composition. The light-yellow area represents the stability fields of the inferred retrograde assemblage. **Figure S9:1 (a)** Isochemical P – T phase diagram calculated for stage M_1 of the kyanite-bearing amphibole-phengite eclogite (#OSR-01B) using the whole-rock bulk composition. The blue dashed area represents the approximate intersection of **(b)** garnet (X_{alm} , X_{prp} and X_{grs}) and rutile (Zr $\mu\text{g/g}$) chemical isopleths, which constraint equilibration at $\sim 590^\circ\text{C}$ – 615°C and 1.7–1.8 GPa. Depicted Zr-rutile isopleths reflect temperature of rutile included in garnet. **Figure S9:2** Isochemical P – T phase diagram calculated for stage M_1 of the kyanite-bearing amphibole-phengite eclogite (#OSR-01B) using the whole-rock bulk composition. The blue area represents the instability fields of the observed retrograde assemblage, defining a P – T range of $\sim 585^\circ\text{C}$ – 615°C and 1.3 GPa. **Figure S10:1 (a)** P – T phase diagram calculated for the eclogitic schist (#OSR-02C) using the corresponding reactive bulk composition for stage M_3 . Grey areas represent the stability field of the inferred assemblage. **Table S11:1** Zircon U–Pb–trace elements data (#RMT-30). **Table S11:2** Zircon U–Pb–trace elements data (G_NIST610). **Table S11:3** Zircon U–Pb–trace elements data (G_NIST612). **Table S12:** Apatite U–Pb–REE data (#RMT-30). **Table S12:1** Garnet Lu–Hf data (#RMT-30). **Table S12:2** Garnet Lu–Hf data (NIST610, GWA-1 and GWA-2). **Table S12:3** Garnet Lu–Hf data (#OSR-01B). **Table S12:4** Garnet Lu–Hf data (NIST610, GWA-1 and GWA-2). **Table S12:5** Garnet Lu–Hf data (#OSR-02C). **Table S12:6** Garnet Lu–Hf data (NIST610, GWA-1 and GWA-2). **Figure S14:1** Cr versus Nb binary diagram (after Meinhold et al., 2008) depicting compositional variation of trace elements in rutile grains from all three samples. The inset is a box-and-whisker plot depicting the Zr contents ($\mu\text{g/g}$) in rutile grains. **Figure S15:1** Geochemical variation of major elements in the eclogites from the Bughea Complex as a function of SiO_2 . Geochemical data for ‘common’ eclogites and the very-high pressure eclogite are from Săbău (2000), Medaris Jr. et al. (2003), Negulescu et al. (2009) and Negulescu and Săbău (2015).