Mesoarchaeon clockwise metamorphic P-T path from the Western Dharwar Craton

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ABSTRACT

In this study, we utilize high-alumina pelitic schist samples from the Holenarsipur Greenstone Belt, a component of the ca. 3.30–3.15 Ga Sargur Supergroup in the Western Dharwar Craton, South India to establish a Mesoarchaeon metamorphic history in the terrane. On the basis of integrated textural, mineral compositional, metamorphic reaction history and P-T pseudosection modelling studies, garnet-monazite trace element thermometry and coupled monazite electron probe micro-analyzer and sensitive high-resolution ion microprobe U-Pb dating results, we constrain for the first time a Mesoarchaeon metamorphism along a hairpin clockwise metamorphic P-T path in the Dharwar Craton. The path reveals a two-stage prograde segment with an initial phase of heating accompanying burial (dP/dT gradient ~ 30 bar/°C) and a later phase of steep burial ($\Delta P \sim 1.1$ kbar, dP/dT gradient ~110 bar/°C), peak metamorphism at 7.5 kbar, 640 °C and a post-peak, steep decompression stage (P~6.3 kbar at 635 °C). Chemically zoned monazite grains shielded within garnet or occurring in low-strain matrix domains were formed in two stages during the Mesoarchaeon metamorphism: (a) Monazite cores with relatively lower Y concentrations, higher Th/U ratios, a strong negative Eu anomaly; their REE partitioning relationships with the garnet inner rim reveal their crystallization was synchronous with the growth of garnet. (b) Monazite rims with higher Y and uraniferous compositions and lower Th/U ratios which, combined with Y and HREE enrichments in the outer rim of garnet, indicate their formation during the post-peak decompression stage associated with coupled decompositions of garnet and zircon. We correlate U-Pb monazite core and rim ages at 3144 ± 7 Ma and 3105 ± 10 Ma respectively with the timings of peak and retrograde metamorphism. Our findings, when evaluated with the available geological, isotopic, geochronological data of Meso- to Palaeoarchaeon extant and extinct felsic crust as well as thermo-mechanical models, predict the ca.3.14–3.11 Ga metamorphism in the Holenarsipur Greenstone Belt as part of a widespread tectono-thermal event in the Western Dharwar Craton. The event marks late stages of sagduction tectonics in the craton as it evolves from a predominantly vertical to a horizontal shortening mode.

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1. Introduction

Divergent opinions exist regarding the timing of initiation of “modern style” plate tectonics in the planet Earth and the indicator materials to be used from natural rock archive (summarised in Korenaga, 2013). Nevertheless, there is a general consensus that the switch over from “pre-subduction”, plume-driven vertical tectonics to the “modern subduction” regime of horizontal plate motions is linked with a change in thermal regime from one of heat generation through radioactive decay to heat loss by conductive cooling (e.g. Herzberg et al., 2010; Michaut and Jaupart, 2007). Stern et al. (2016) propose that this tectonic transition took place in the time period between 3.2 and 2.8 Ga.

Metamorphic rocks and well-constrained metamorphic P-T-t paths have emerged as potential tools to model Mesoarchaeon to Neoarchaeon fossil tectonic settings (Cutts et al., 2014; Dziggel et al., 2002; François et al., 2014; Li et al., 2018; Moyen et al., 2006; Nicoli et al., 2015; Roberts and Santosh, 2018; Sizova et al., 2018; Van Kranendonk, 2011a). However, despite its potential significance, well-preserved amphibolite-granulite facies Mesoarchaeon crustal sections are rather rare in the global rock record, in part due to their incomplete preservation and also due to the effects of strong superposed Neoarchaeon deformation and metamorphism that, in most cases, have obliterated the older history. One of the key challenges in applying metamorphic constraints to model the Meso- to Neoarchaeon plate tectonic transition is to look through the veil of superposed orogenesis, and obtain Mesoarchaeon time information in a metamorphic context. Monazite petrochronology that links monazite ages in a metamorphic context is

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particularly suitable in this context (Bhowmik et al., 2012, 2014; Engi et al., 2017; Foster et al., 2004; Gibson et al., 2004; Hermann and Rubatto, 2003; Kelsey et al., 2008; Kohn and Malloy, 2004; Montel et al., 1996; Pyle et al., 2001; Rubatto et al., 2001, 2013; Williams et al., 2007).

In this study, we address this issue using rocks from a Mesoarchaeoan, Sargur-type greenstone belt, locally referred to as the Holenarsipur Greenstone Belt (HGB) in the Western Dharwar Craton (WDC), South India. Although, previous studies have suggested a major tectono-thermal and tectono-magmatic event in the WDC that led to the cratonization of the Dharwarian crust at ca. 3.0 Ga (Jayananda et al., 2013; Jayananda et al., 2015; Raith et al., 1982), the metamorphic conditions and P-T path of evolution of this Mesoarchaeoan event are yet to be established. On the other hand, preliminary geochronological investigations in the HGB and other segments of the Sargur supracrustals have recorded a pervasive latest Neoarchaeoan-earliest Palaeoproterozoic regional metamorphism in the WDC (Bouhallier, 1995; Jayananda et al., 2013). In order to resolve this issue of Mesoarchaeoan vs. Neoarchaeoan-earliest Palaeoproterozoic orogenesis in the WDC, we have adopted an integrated approach utilising reconstruction of the metamorphic P-T path through phase equilibria modelling, garnet-monzonite thermometry and coupled Electron Probe MicroAnalysis (EPMA) and sensitive High-Resolution Ion Microprobe (SHRIMP) U-Pb monazite dating techniques. These are applied to a group of middle-amphibolite facies pelitic schist samples from the HGB. Combining trace element analyses of garnet and monazite, in situ Y, Th, and U chemical mapping of compositionally zoned and texturally well constrained monazite grains, results of chemical and SHRIMP U-Pb dating of same compositional domains of monazite and metamorphic P-T path constraint, we establish a Mesoarchaeoan

metamorphism in the WDC. We discuss the geodynamic significance of this metamorphic event in the overall context of plate tectonic framework at this time period.

2. Geological setting

The WDC is located in the southern, cratonic part of the Indian Peninsula (Fig. 1a) and consists mainly of large swaths of tonalite-trondhjemite-granodioritic (TTG) gneiss (locally known as the Peninsular Gneiss complex) with intercalations of linear, generally N-S trending, belts of metamorphosed volcano-sedimentary units and associated granite greenstone belts (Swami Nath and Ramakrishnan, 1981). The TTG gneisses record three periods of felsic magmatism in the terrane at 3411–3276 Ma, 3230–3200 Ma and 3134–3103 Ma (Beckinsale et al., 1980, 1982; Bhaskar Rao et al., 1983, 2008; Guitreau et al., 2017; Jayananda et al., 2015; Peucat et al., 1993). Existence of an even older proto-crust up to 3610 Ma in the WDC is indicated by published detrital and inherited U-Pb zircon ages in metasedimentary rocks and TTG gneisses (Guitreau et al., 2017; Hokada et al., 2013; Nutman et al., 1992; Sarma et al., 2012).

The volcano-sedimentary units are broadly subdivided into two types: the older (c.3.3–3.1 Ga) Sargur Supergroup and younger (2.91–2.56 Ga) Dharwar Supergroup, exposed as separate linear belts (Fig. 1a). The focus of this investigation is the older Sargur belt rocks (Fig. 1a) that are dominated by metamorphosed komatiite-basalt volcanics with rare felsic volcanics and interlayered sedimentary rocks (Swami Nath and Ramakrishnan, 1981). The antiquity of the Sargur supracrustals is indicated by the following observations: (a) Sm-Nd whole-rock isochron age of komatiite between 3352 ± 110 Ma (Jayananda et al., 2008) and 3125 ± 120 Ma (Mukherjee et al., 2012)

Fig. 1. (a) Simplified geological map of the Western Dharwar Craton and the adjoining Coorg granulite massif. Published Mesoarchaeoan geochronological constraints from different parts of the craton are listed below. Age (in Ma) with superscript as a reference to authors: 1, Peucat et al. (1995); 2, Jayananda et al. (2008); 3, Mukherjee et al. (2012); 4, Jayananda et al. (2015); 5, Guitreau et al. (2017). Age with subscript as dating methods. 1, U-Pb Zrn; 2, U-Pb Ttn; 3, Sm-Nd WR. Rectangular box shows the location of the Holenarsipur Greenstone Belt, shown in detail in Fig. 1b. Inset shows the two fold subdivision of the Dharwar craton into western (WDC) and eastern (EDC) types, being separated by the Closepet Granite. (b) Simplified geological map of the Holenarsipur Greenstone Belt (HGB) and adjoining Halekote trondhjemite pluton and Garur TTG gneisses. Also shown are the locations of published geochronological data and the samples of this study. Age (in Ga) with superscript: Author reference; 1, Guitreau et al. (2017); 2, Jayananda et al. (2015). Age with subscript: Dating method; 1, SHRIMP U-Pb Zrn; 2, SHRIMP U-Pb Ttn.

for the timing of eruption of komatiite lavas, (b) 3298 ± 7 Ma U-Pb zircon age of rhyolite volcanism in the HGB (Peucat et al., 1995), (c) detrital zircons in associated sedimentary rocks, from different parts of the Sargur belt ranging in age from 3580 to 2960 Ma (Bhaskar Rao et al., 2008; Maibam et al., 2011; Nutman et al., 1992) and (d) intrusive granitoids in the Sargurs between 3.2 and 3.1 Ga (Beckinsale et al., 1982; Guitreau et al., 2017; Jayananda et al., 2015; Peucat et al., 1993).

The metamorphic grade of the Sargur supracrustals increases from middle-amphibolite grade in the north to granulite grade in the southern-most parts of these belts (Janardhan et al., 1978, 1982; Jayananda et al., 2013; Raase et al., 1986; Rollinson et al., 1981; Srikantappa et al., 1985) (Fig. 1a). There is considerable debate regarding the timescales of metamorphism of the Sargur supracrustals and the WDC. Garnet Sm-Nd dating of medium-grade (Holenarsipur domain) and high-grade (type Sargur and adjoining Gundlupet domains) Sargurs (Fig. 1a) have yielded a uniform garnet growth age between 2.53 and 2.43 Ga (Bouhallier, 1995; Jayananda et al., 2013). This signifies a latest Neoproterozoic-earliest Palaeoproterozoic regional metamorphic event in the WDC. In contrast, imprints of an older Mesoarchaeon tectono-thermal and tectono-magmatic event in the WDC have been inferred based on (a) resetting of Pb isotope ratios of detrital zircon in the Sargur supracrustals between 3.1 and 3.0 Ga (Lancaster et al., 2015; Nutman et al., 1992), (b) crustal anatexis and formation of a grey granite vein at 3.15 Ga within older ca. 3.35 Ga Garur gneiss (Jayananda et al., 2015), (c) development of zircon overgrowth at 2.99 Ga in 3.2 Ga potassic granites (Jayananda et al., 2013) and (d) metamorphic recrystallization of 3.2 Ga Halekote-equivalent trondhjemitic gneiss at 3.1 Ga (Guitreau et al., 2017).

We have chosen the Holenarsipur Greenstone Belt (HGB), the largest preserved exposure of the Sargur supracrustal unit, to find out the possible signatures of an older Mesoarchaeon tectono-thermal event, despite an overprint of a strong Neoproterozoic-Palaeoproterozoic regional event. This is because (a) the HGB and surrounding TTG gneisses constitute the oldest nucleus of the Dharwar craton (e.g. Radhakrishna and Naqvi, 1986), (b) previous studies have established an intrusive relationship between the Mesoarchaeon trondhjemitic pluton (e.g. Halekote pluton) and the Sargur supracrustals (e.g. Ramakrishnan and Viswanatha, 1981), (c) the dome and keel Garur gneiss-trondhjemitic and greenstone outcrop pattern has been previously explained by a model of vertical tectonics of diapiric domes and sagducted greenstones (Bouhallier, 1995; Bouhallier et al., 1993, 1995). The timing of the vertical tectonics is, however, debatable and has been inferred to be of either Mesoarchaeon (Bouhallier et al., 1993) or Neoproterozoic (Bouhallier, 1995) age, and (d) pelitic schists of the HGB show the occurrences of chloritoid, garnet, staurolite and kyanite with plagioclase, two micas, ilmenite, rutile, quartz and accessory monazite (Bouhallier, 1995; Bouhallier et al., 1993; Bouhallier et al., 1995; Jayananda et al., 2013; Swami Nath and Ramakrishnan, 1981).

Bouhallier (1995) previously studied the metamorphic evolution of four distinct types of metapelitic schists (e.g. garnet-chloritoid schist, garnet-staurolite-epidote schist, staurolite-kyanite schist and garnet-staurolite-kyanite schist) from different strain domains in the HGB. The study revealed a broadly synkinematic growth of garnet with or without chloritoid, staurolite and kyanite at metamorphic peak and the same peak P-T estimates at 7–8 kbar, 600 °C at linear troughs and foliation triple points in the greenstone. The foliation triple points have additionally recorded a prograde burial history from 3 kbar, 470 °C to 7–8 kbar, 600 °C, which was taken as evidence for the “diapiric instabilities model”. Bouhallier (1995) suggested a ca. 2.5 Ga age of the metamorphism in the HGB. Summarising, the mineralogy and reaction textures in the Holenarsipur pelitic schists offer excellent opportunities to reconstruct the metamorphic P-T path of evolution, obtain texturally-controlled monazite dates and interpret monazite dates in a metamorphic context.

Given this context of a strong regional scale ca. 2.5 Ga tectono-thermal event affecting the HGB, we suggest that imprints of the older Mesoarchaeon event are likely to be preserved in low-strain domains with respect to the last pervasive deformation event. In order to identify such samples with records of an earlier metamorphic history, we have carried out studies of metamorphism of 8 pelitic schist samples, spread over 5 locations in different parts from the southern margin of the HGB (sample locations in Fig. 1b). The pelitic schists, which are interlayered with fuchsite quartzite, ultramafic schist and amphibolite show a pervasive biotite + muscovite + chlorite $\pm$ staurolite-defined metamorphic foliation. On the basis of careful textural analysis, we have classified them into two groups: (a) Samples recording syn- to post-kinematic growth of porphyroblasts (e.g. staurolite, garnet and kyanite) with respect to the foliation stated above, and similar to that recorded by Bouhallier (1995). Pelitic schists of this type (e.g. samples HNP-1, HNP-1C, 09-HN-01, HNA-1A, HG-6B, HG-8) have undergone peak kyanite-, and locally kyanite + garnet or staurolite-zone metamorphism, generally in the stability field of rutile. (b) In the second group, pelitic schists are polyphase metamorphosed with porphyroblasts having grown both pre- and post-kinematic with respect to the foliation (samples HNP-10 and 09-HNP-10). Although, this second group of polyphase metamorphosed samples are less abundant, these are potentially significant to constrain an earlier phase of metamorphism. For this reason, we have selected samples HNP-10 and 09-HNP-10 from two nearby outcrops in the same location, and also in close spatial association with the Halekote trondhjemitic pluton (Fig. 1b) to obtain metamorphic P-T path information of the Mesoarchaeon event, the primary focus of this study. We shall present the metamorphic P-T-t path information of the second group of samples that have preferentially registered the younger event in a separate publication.

3. Methods of study

3.1. Analytical techniques

3.1.1. Electron Probe Micro-Analyzer (EPMA)

Major-element mineral chemical compositions were measured using a CAMECA SX-100 electron microprobe fitted with four spectrometers at the Department of Geology and Geophysics, Indian Institute of Technology (IIT), Kharagpur. Analyses were performed with an accelerating voltage of 15 kV, a beam current of 15 nA and a beam size of $\sim 1 \mu\text{m}$ ($\sim 5 \mu\text{m}$ for feldspar and micas). Natural mineral standards were used for calibration, including orthoclase (for K and Si), jadeite (Na), wollastonite (Ca), periclase (Mg), hematite (Fe), corundum (Al), rhodonite (Mn) and chromite (Cr). Analytical variation within standards was kept at or below ~ 1.5 wt% for primary standards. Representative mineral chemical analyses (presented in Table 1) were recalculated using the AX program (THERMOCALC software, v. 3.26).

For EPMA dating, monazites were analysed with the same EPMA at an accelerating voltage of 20 kV, a beam current of 150 nA, and a beam size of $1 \mu\text{m}$. Spots for monazite analyses were selected on the basis of Y-, Th- and U- X-ray element maps. These were generated at 20 kV and 50 nA, with 50–75 ms dwell times and step sizes at 0.2–0.7 μm . X-ray data were calibrated against natural and synthetic standards. These were: yttrium aluminium garnet (YAG) for Y; ThO₂ for Th; UO₂ for U; vanadinite for Pb; synthetic silica-aluminium glass containing 4% REE for La, Ce, Nd, Pr, Sm, Ho, Dy and Gd; apatite for P and Ca; corundum for Al; hematite for Fe; Th-glass for Si.

The measured X-ray lines for different elements and the attendant crystals were: Th M α , U M β and Ca K α (PET); Pb M α , La-Ce L α and P K α (LPET); Pr-Nd-Gd-Ho L β , Sm-Dy L α and Fe K α (LIF); Y L α , Al-Si K α (TAP). The counting times for these elements on the peak were: 10s for Si, Al, Fe and P; 20s for Ca, La and Ce; 40s for Nd, Sm, Gd and Y; 60s for Dy and Ho; 200 s for Th and U and 300 s for Pb. Respective background intensities were measured on both sides of the peak for

Table 1

Representative mineral chemical analyses in pelitic schists from the Holenarsipur Greenstone Belt, Western Dharwar Craton, India.

Sample no.	HNP-10-1	HNP-10-2	HNP-10-1	HNP-10-1	HNP-10-2	HNP-10-1	HNP-10-1	HNP-10-1	HNP-10-1	HNP-10-2	HNP-10-1	HNP-10-1	HNP-10-1	HNP-10-1	HNP-10-1	HNP-10-2
Spot no.	69_1	15_2	69_37	69_40	10_1	36_1	57_1	1_1	38_1	12_1	30_1	18_1	45_1	14_1	11_1	22_1
Mineral	Grt_A	Grt_E	Grt_A	Grt_A	Grt_E	Bt	Bt	Bt	Bt	Bt	Bt	Chl	Chl	Chl	St	St
Site	P(C)	P(IR)	P(OR)	P(E)	P(OR)	M(FOL)	M(FOL)	I ⁺ Grt	I ⁺ Grt	Cor > Grt	Cor > Grt	Cor > Grt	I ⁺ Grt	FOL	M(FOL)	M(L)
SiO ₂	37.31	36.83	36.06	36.14	36.91	36.53	36.43	37.02	36.86	36.53	36.86	25.28	24.7	25.12	27.54	27.64
TiO ₂	0.01	b.d.l.	b.d.l.	b.d.l.	b.d.l.	1.67	1.48	1.47	1.55	2.17	1.49	0.08	b.d.l.	0.06	0.57	0.57
Al ₂ O ₃	21.30	21.32	21.07	20.96	21.41	19.05	18.93	19.34	19.48	18.83	19.31	23.89	22	23.76	53.15	52.86
Cr ₂ O ₃	0.03	b.d.l.	0.04	0.04	b.d.l.	0.01	0.02	0.14	0.03	0.14	0.11	0.07	0.02	0.02	0.21	0.29
FeO	31.72	34.41	33.52	34.2	34.67	16.87	16.09	15.76	16.33	16.93	16.12	20.52	26.16	21.17	13.74	14.57
MnO	5.22	3.3	3.42	3.88	3.63	0.12	0.04	0.09	b.d.l.	0.05	b.d.l.	0.03	0.21	0.14	0.14	0.15
MgO	3.52	3.96	3.78	3.15	3.24	12.29	12.66	13.2	12.4	12.06	12.19	18.34	14.57	18.47	1.91	1.98
CaO	0.65	0.79	1.27	1.09	1.31	0.02	b.d.l.	0.02	b.d.l.	0.01	0.12	0.04	0.01	0.06	0.03	0.01
Na ₂ O	0.03	b.d.l.	0.02	0.02	b.d.l.	0.38	0.38	0.43	0.4	0.27	0.39	0.02	0.02	0.01	b.d.l.	0.02
K ₂ O	b.d.l.	b.d.l.	0.01	0.03	b.d.l.	8.99	8.74	8.63	8.79	8.67	7.67	b.d.l.	b.d.l.	0.02	b.d.l.	0.01
Total	99.79	100.61	99.19	99.5	101.16	95.94	94.78	96.10	95.83	95.66	94.27	88.27	87.69	88.82	97.29	98.1
Oxygen	12	12	12	12	12	11	11	11	11	11	11	14	14	14	46	46
Si	3.00	2.94	2.92	2.93	2.94	2.72	2.72	2.72	2.73	2.72	2.75	2.57	2.62	2.55	7.69	7.69
Ti	0.00	–	–	–	–	0.09	0.08	–	0.08	0.12	0.08	0.01	–	0.01	0.12	0.12
Al	2.02	2.01	2.01	2.01	2.01	1.67	1.67	1.68	1.7	1.65	1.70	1.75	2.75	2.84	17.50	17.35
Cr	0.00	–	0.00	0.00	b.d.l.	b.d.l.	0.00	0.01	0.00	0.01	0.01	0.00	0.00	0.00	0.05	0.06
Fe ²⁺	2.13	2.29	2.26	2.31	2.30	1.05	1.00	0.96	1.01	1.05	1.01	1.75	2.31	1.80	3.21	3.39
Mn	0.36	0.22	0.23	0.27	0.25	0.01	0.00	0.01	–	0.00	–	0.00	0.02	0.01	0.03	0.04
Mg	0.42	0.47	0.46	0.38	0.39	1.36	1.41	1.45	1.37	1.34	1.36	2.78	2.3	2.79	0.79	0.82
Ca	0.06	0.07	0.11	0.10	0.11	0.00	b.d.l.	0.00	–	0.00	0.01	0.00	0.00	0.01	0.01	0.00
Na	0.00	–	–	0.00	–	0.05	0.05	0.06	0.06	0.04	0.06	0.00	0.00	0.00	–	0.01
K	–	–	–	0.00	–	0.85	0.83	0.81	0.83	0.82	0.73	0.00	–	0.00	–	0.00
Sum	8.00	8.00	8.00	8.00	8.00	7.81	7.79	7.77	7.78	7.76	7.70	9.99	10	10	29.41	29.49
X _{Mg}	0.16	0.17	0.17	0.14	0.14	0.57	0.59	0.60	0.58	0.56	0.57	0.61	0.50	0.61	0.20	0.19
X _{Prp}	0.14	0.16	0.15	0.12	0.13	–	–	–	–	–	–	–	–	–	–	–
X _{Grs}	0.02	0.02	0.04	0.03	0.04	–	–	–	–	–	–	–	–	–	–	–
X _{Alm}	0.72	0.75	0.74	0.76	0.76	–	–	–	–	–	–	–	–	–	–	–
X _{Sps}	0.12	0.07	0.08	0.09	0.08	–	–	–	–	–	–	–	–	–	–	–

Sample no.	HNP-10-1	HNP-10-1	HNP-10-2	HNP-10-2	HNP-10-2	HNP-10-2	HNP-10-1	HNP-10-2	HNP-10-2	HNP-10-2	HNP-10-1	Sample no.	HNP-10-1
Spot no.	16_1	53_1	1_1	14_1	8_1	10_1	73_1	28_1	18_1	59_1	55_1	Spot no.	1_3
Mineral	Ms	Ms	Pa	Pl	Pl	Pl	Pl	Pl	Pl	Ilm	Sil	Mineral	Ap
Site	M	M	I ⁺ Grt	I ⁺ Grt	M(C)	M(C)	M(R)	M(R)	M(R)	M	I ⁺ Ms	Site	I ⁺ Grt
SiO ₂	45.64	47.25	49.88	65.21	64.39	64.08	63.07	62.2	62.87	0.03	37.09	SiO ₂	0.03
TiO ₂	0.31	0.31	0.04	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	52.85	b.d.l.	TiO ₂	0.02
Al ₂ O ₃	36.42	37.19	38.68	21.11	21.82	22.18	22.81	22.57	22.97	0.20	62.73	Al ₂ O ₃	0.01
Cr ₂ O ₃	0.12	0.14	0.15	0.03	0.01	0.01	b.d.l.	b.d.l.	b.d.l.	0.01	0.01	FeO	0.57
FeO	1.20	0.73	1.9	0.75	b.d.l.	b.d.l.	b.d.l.	0.08	0.06	44.45	0.04	MnO	0.10
MnO	b.d.l.	0.02	0.12	0.06	b.d.l.	0.12	0.01	0.05	0.05	1.30	b.d.l.	CaO	56.35
MgO	0.41	0.42	0.32	b.d.l.	b.d.l.	0.03	0.01	0.01	0.02	0.08	b.d.l.	K ₂ O	0.10
CaO	b.d.l.	0.01	0.12	2.41	3.23	3.32	4.3	4.64	4.54	0.01	b.d.l.	P ₂ O ₅	41.32
Na ₂ O	2.24	2.10	5.12	10.72	10.19	9.9	9.57	9.08	9.39	b.d.l.	0.01	F	3.64
K ₂ O	8.10	8.44	2.65	0.02	0.04	0.07	0.05	0.05	0.05	b.d.l.	0.04	Cl	0.09
Total	94.43	96.61	98.98	100.32	99.68	99.7	99.81	98.69	99.95	98.93	99.93	Total	102.14

(continued on next page)

Table 1 (continued)

Sample no.	HNP-10-1	HNP-10-1	HNP-10-2	HNP-10-2	HNP-10-2	HNP-10-2	HNP-10-2	HNP-10-1	HNP-10-2	HNP-10-2	HNP-10-2	HNP-10-2	HNP-10-1	Sample no.	HNP-10-1
Oxygen	11	11	8	8	8	8	8	8	8	8	3	5	-O = F, Cl	1.55	
Si	3.04	3.07	2.87	2.85	2.84	2.80	2.79	2.79	2.79	2.79	0.00	1.00	Total _{Rec}	100.58	
Ti	0.02	0.02	-	-	-	-	-	-	-	-	1.01	-	Oxygen	25	
Al	2.86	2.84	1.10	1.14	1.16	1.19	1.19	1.19	1.2	1.2	0.01	2.00	Si	0.01	
Cr	0.01	0.01	0.00	0.00	0.00	-	-	-	-	-	0.00	0.00	Ti	0.00	
Fe ²⁺	0.07	0.04	0.03	-	-	-	0.00	0.00	0.00	0.00	0.94	0.00	Al	0.00	
Mn	-	0.00	0.00	-	0.00	0.00	0.00	0.00	0.00	0.00	0.03	-	Fe ²⁺	0.08	
Mg	0.04	0.04	0.03	-	-	-	-	-	-	-	0.00	-	Mn	0.01	
Ca	-	0.00	0.11	0.15	0.16	0.20	0.21	0.21	0.22	0.22	0.00	-	Ca	9.78	
Na	0.29	0.26	0.92	0.88	0.85	0.82	0.79	0.79	0.81	0.81	0.00	0.00	K	0.00	
K	0.69	0.70	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	P	5.67	
Sum	7.00	6.98	5.03	5.02	5.01	5.02	5.01	5.01	5.02	5.02	1.99	3.00	F	1.86	
X _{Ms}	0.58	0.62	-	-	-	-	-	-	-	-	-	-	Cl	0.02	
X _{Pa}	0.29	0.26	0.62	-	-	-	-	-	-	-	-	-	X _F	0.93	
X _{Cel}	0.11	0.08	-	-	-	-	-	-	-	-	-	-	X _{Cl}	0.01	
X _{Prl}	0.02	0.04	-	-	-	-	-	0.21	0.22	0.22	-	-	X _{OH}	0.06	
X _{An}	-	-	0.11	0.15	0.16	0.20	0.21	-	-	-	-	-	-	-	
X _{Alm}	-	-	-	-	-	-	-	-	-	-	0.97	-	-	-	
X _{Pyrophanite}	-	-	-	-	-	-	-	-	-	-	0.03	-	-	-	

Mineral abbreviations used after Kretz (1983); Other abbreviations used: P, Porphyroblast; C/R/IR/OR/E/M/FOL, Core/Rim/Inner Rim/Outer Rim/Edge/Matrix/Foliation; L//COR: Late Inclusion/Corona; ^/>: Within/Against; X_{Mg} [= Mg/(Mg + Fe²⁺)]

half the peak times. Interference of Y L α on Pb M α , Th M ζ on Pb M α and Th M γ on U M β were corrected after measuring the interfering lines during calibration and thereafter applying the overlap correction. Detection limits were 155 ppm for Y, 145 ppm for Pb, 380 ppm for Th and 300 ppm for U. Further details of the monazite analytical protocol were presented in [Bhowmik et al. \(2014\)](#).

Prior to and during analytical sessions, analyses of an in-house reference monazite age standard (thermal ionization mass spectrometry age at 483 ± 1 Ma, Chatterjee et al., 2010) were carried out periodically to monitor short-term systematic errors and to check for overall accuracy. This has yielded a long-term weighted mean age of 478 ± 4 Ma (2σ error, MSWD = 0.57, $n = 85$). Analyses of the monazite standard in two analytical sessions of this study yielded a mean age of 493 ± 16 Ma (MSWD = 0.54, $n = 6$), which is within 3% of its long-term average. Relative age probability plots and weighted mean ages of monazite were computed with the Isoplot program (Ludwig, 2003). The complete monazite compositional dataset has been presented in the Supplementary Table S1.

3.1.2. Laser ablation-inductively coupled plasma mass spectrometry (LA-ICPMS)

Trace element measurements in garnet and monazite were carried out at the Department of Geology and Geophysics, IIT, Kharagpur using a Thermo-Fisher Scientific iCap-Q quadrupole-inductively coupled plasma mass spectrometer (ICPMS) attached to a new wave 193 Ar-F Excimer laser ablation system. The ablation was done at 5 Hz pulse frequency, 5 J/cm² beam energy density and spot sizes ranged from 30 to 40 μm depending on the grain size. The analyses were performed in time resolved mode with each analysis consisting of 30s background measurement and 40s peak signal measurement. The ICPMS was optimized for maximum sensitivity on Li, Co, In, Th, and U using the NIST 612 reference glass. Calibration was done using NIST 610 and NIST 612 glass standards. Precision as determined from repeat analyses of the NIST 612 and NIST 610 glasses are better than 5% (2σ) and 11% (2σ) respectively. Representative trace element analyses of garnet and monazite are presented in [Table 2](#).

3.1.3. SHRIMP U-Pb monazite dating

The same monazite grains that were dated by the EPMA were also analysed with the Sensitive High Resolution Ion Micro-Probe Reverse Geometry (SHRIMP RG) housed at the Research School of Earth Sciences (RSES), Australian National University, Canberra. Regions of interest (8 mm diameter) with selected monazite and associated minerals were removed from standard polished thin sections with an ultrasonic disk drill. Grains of the Wendell monazite reference material (R. Tucker *pers. commun.*) were mounted in epoxy and polished; sections of the standard were then cut from the epoxy mount. The cut thin sections were then mounted on double-sided tape together with the reference standards and cast in 25 mm diameter epoxy mounts. The final mounts were then coated with gold to prevent sample charging during sputtering (Ireland and Williams, 2003).

The operating conditions and techniques for SHRIMP U-Th-Pb analysis of monazite followed that described by Ireland and Gibson (1998). The primary ion beam size was 30 μm diameter, with an intensity of ca. 4 nA. Mass resolution was set to 5000 $M/\Delta M$ (full width at 10% peak height), which is sufficient to resolve most significant molecular interferences. However, a potential isobaric interference from monazite (ThNdO_2^+ at mass 204) cannot be resolved from $^{204}\text{Pb}^+$ (Ireland et al., 1999). Energy filtering discriminates against polyatomic molecules that are tightly constrained at the extraction energy (10 keV), whilst maintaining the presence of atomic and oxide species with long energy distributions (Ireland, 1995). For this work we applied an energy offset of approximately 15 eV which effectively removes the ThNdO_2^+ interference from $^{204}\text{Pb}^+$.

Isotopic ratios were measured by repeated cycles through the relevant molecular and nuclidic masses: 203 ($[^{140}\text{Ce}^{31}\text{P}^{16}\text{O}_2]^+$),

204 ($^{204}\text{Pb}^+$), 204.1 (background), 206 ($^{206}\text{Pb}^+$), 207 ($^{207}\text{Pb}^+$), 208 ($^{208}\text{Pb}^+$), 232 ($^{232}\text{Th}^+$), 238 ($^{238}\text{U}^+$), 248 ($^{232}\text{Th}^{16}\text{O}^+$) and 254 ($^{238}\text{U}^{16}\text{O}^+$). The 300 Ma Wendell standard ($^{206}\text{Pb}/^{238}\text{U}$ age analysed by P. Holden, *pers. commun.*) was used to normalize measured $^{206}\text{Pb}^+/\text{U}^+$ and $^{208}\text{Pb}^+/\text{Th}^+$ ratios, with independent power-law calibrations for $^{206}\text{Pb}^+/\text{U}^+$ [v. (UO^+/U^+)] and $^{208}\text{Pb}^+/\text{Th}^+$ [v. (ThO^+/Th^+)] (Ireland and Gibson, 1998). This was accommodated through data reduction with the well established ANU in-house software PRAWN and LLEAD. Data were compared to output from POXI, an in-house ANU LabView program that includes the data reduction methods of SQUID 1 (Ludwig, 2001), as well as PRAWN. For the purposes of this work, PRAWN/LLEAD was favoured because the full U-Pb, Pb-Pb, and Th-Pb assessment could be made in a consistent and established package (Ireland and Gibson, 1998).

All data have been corrected for common Pb on the basis of the $^{204}\text{Pb}/^{206}\text{Pb}$ ratio. Corrections for common Pb are relatively small with all unknown grains having <0.2% common ^{206}Pb . Calculated 1σ external spot to spot uncertainties for the Wendell standard for $^{206}\text{Pb}/^{238}\text{U}$ is 2.3%, and that for $^{208}\text{Pb}/^{232}\text{Th}$ is 2.0%. These uncertainties were combined with the analytical uncertainties of individual measurements. The reported ages are >3000 Ma and so the $^{207}\text{Pb}/^{206}\text{Pb}$ chronometer offers the highest precision (Table 4). Single analyses in the data table and concordia plot are presented with 1σ errors, whereas uncertainties in weighted mean ages are quoted at 2σ , unless otherwise indicated.

3.2. Metamorphic reconstruction

3.2.1. P-T Pseudosection modelling

We quantify the mineral reaction pathways of the Mesoarchaean event using a calculated metamorphic phase diagram. We have adopted an approach of calculation of an equilibrium phase diagram using a Gibbs free energy minimization technique in P-T space (pseudosections) for fixed bulk rock compositions (Connolly, 1990). P-T pseudosections have been computed in the model system $\text{MnO} - \text{TiO}_2 - \text{Na}_2\text{O} - \text{CaO} - \text{K}_2\text{O} - \text{FeO} - \text{MgO} - \text{Al}_2\text{O}_3 - \text{SiO}_2 - \text{H}_2\text{O}$ (MnTiNCKFMASH) and using the PERPLEX software (v. 6.7.3., updated at June 8, 2016) and the thermodynamic dataset of Holland and Powell (1998). The bulk rock composition has been adjusted for Ca by considering its amount stored in apatite. The following mixing models were applied: garnet: Ganguly et al. (1996); plagioclase: Newton et al. (1980); biotite: Powell and Holland (1999); chlorite: Holland and Powell (1998); chloritoid/staurolite: Holland and Powell (1998); white mica: Smye et al. (2010); ilmenite: Ideal mixing; cordierite:

Ideal mixing; alkali feldspar: Waldbaum and Thompson (1968). Kyanite, sillimanite, rutile and quartz are considered as pure phases. The metamorphic phase relationships for quartz and water saturated condition are presented in Fig. 4.

3.2.2. Garnet-monazite thermometry

We have additionally carried out garnet-monazite thermometry following the calibration of Pyle et al. (2001) to estimate peak metamorphic temperature and compare the result with that obtained by pseudosection modelling. Application of the thermometer requires compositional information of the phases involved in the chemically balanced reaction: $\text{YAG} (\text{Y} + \text{Al Garnet}) + (\text{OH})\text{-Apatite} + (25/4) \text{ Quartz} = (5/4) \text{ Grossular} + (5/4) \text{ Anorthite} + 3 \text{ YPO}_4 (\text{Monazite}) + (1/2) \text{ H}_2\text{O}$. The methodology is shown to have yielded meaningful temperature estimates in medium- and high-grade pelitic schists and migmatites respectively (Hallett and Spear, 2015; Pyle et al., 2001).

4. Petrography and mineral chemistry

4.1. Petrography

4.1.1. Sample HNP-10

This sample is a high alumina, garnet - staurolite - sillimanite - biotite - muscovite - chlorite - ilmenite - plagioclase - quartz schist with minor paragonite, monazite and zircon. We have studied two slices of the sample (named HNP-10-1 and HNP-10-2). Both the slices show a low abundance of garnet (1.5–3 modal%) and are thinly banded with alternate layers rich in staurolite + ilmenite $\pm$ chlorite, biotite + muscovite $\pm$ chlorite and quartz + plagioclase, defining a pervasive foliation (Fig. 2a–b). The foliation wraps around garnet porphyroblasts (Fig. 2c). In the interfolial domain, there are coarse aggregates of staurolite (Fig. 2d) and locally muscovite (Fig. 2e). We consider these staurolite and muscovite aggregates to have pre-dated the pervasive foliation. The foliation is also overgrown by a later generation biotite (Fig. 2a), staurolite (Fig. 2f) and chlorite (Fig. 2g). Garnet is generally inclusion-free (Fig. 2c), excepting minor presence of biotite, paragonite (Fig. 2h), quartz and chlorite (Fig. 2i) within it. It is partially rimmed by biotite (Fig. 2a,i). Coarse matrix muscovite grains in the interfolial domain show presence of thick seams of fibrolitic sillimanite, particularly in its core (Fig. 2e). Thin flakes of muscovite in the foliation domain are characteristically free of sillimanite (Fig. 2j). In HNP-10, the analysed monazite grains occur in three textural settings (Fig. 2a): inclusions within garnet (grains M1 to M3 in HNP-10-1 and M1 in HNP-10-2), inclusions

Table 2

Representative trace element analyses of garnet and monazite in the pelitic schist sample HNP-10-1.

Mineral	Grt	Grt	Grt	Grt	Grt	Grt	Mnz	Mnz	Mnz	Mnz
Spot No.	1	2	3	4	6	8	5	7	9	10
Site	P(IC)	P(OC)	P(IR)	P(OR)	P(E)	P(E)	C	C	C	C
Y	198.4	178.0	125.4	88.9	111.2	116.9	11,237	10,534	11,371	10,253
La	b.d.l.	b.d.l.	b.d.l.	b.d.l.	0.5	2.3	115,611	114,733	127,049	98,285.05
Ce	b.d.l.	b.d.l.	b.d.l.	b.d.l.	1.2	6.6	189,016	184,072	202,442	164,645
Pr	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	0.2	21,169	20,593	22,577	18,083
Nd	b.d.l.	0.10	<0.15	b.d.l.	0.1	3.3	83,649	78,642	89,857	74,382
Sm	0.3	0.1	0.1	0.1	0.1	0.3	15,151	14,505	15,350	13,286
Eu	0.2	0.1	0.1	0.1	0.2	0.2	2018	1911	2301	1803
Gd	2.1	2.2	0.9	0.6	0.5	1.5	12,375	11,557	11,681	10,323
Tb	1.5	1.6	0.9	0.6	0.5	0.6	1420	1292	1316	1236
Dy	25.9	23.2	15.2	8.9	11.1	11.0	4916	4600	4568	4377
Ho	8.3	7.9	5.7	3.8	4.9	5.0	486.7	447.2	461.2	421.3
Er	27.8	27.2	21.5	18.3	25.5	25.6	565.9	531.2	564.7	472.2
Tm	4.8	4.7	3.5	4.4	5.5	5.8	31.8	29.3	31.4	23.4
Yb	35.7	32.6	28.2	41.5	55.1	58.8	102.1	96.9	115.9	75.7
Lu	4.1	3.9	3.1	6.7	9.7	10.3	8.1	7.4	9.0	6.4
(HREE) _{Tot}	110.2	103.4	78.9	84.9	112.7	118.5	19,905	18,561	18,747	16,935
Lu/Dy	0.16	0.17	0.21	0.76	0.88	0.94	0.003	0.002	0.002	0.002

Element concentrations in ppm; (HREE)_{Tot} = (Gd + Tb + Dy + Ho + Er + Tm + Yb + Lu).

within quartz in the interfolial domain (grain M5 in HNP-10-1) and in the foliation domain (grains M4, M6, M7 in HNP-10-1 and M2 to M4 in HNP-10-2). Whereas in the first two textural settings, shapes of monazite grains vary from partially elongated (Fig. 5a–b) through rectangular, ellipsoidal to ovoid (Fig. 5d–e), these are predominantly elongated in the foliation domain (Fig. 5g–h, j–k).

4.1.2. Sample 09-HNP-10

The high alumina pelite sample 09-HNP-10 is a staurolite - biotite - muscovite - quartz schist, and differs from HNP-10 by the following textural attributes: (a) presence of kyanite (Fig. 2k) and rutile (Fig. 2l), (b) absence of garnet, sillimanite, paragonite and ilmenite and (c) absence of mineralogical layering. The matrix foliation is defined by oriented muscovite, biotite, staurolite and rarely chlorite (Fig. 2m). Porphyroblasts of kyanite overgrow the foliation. There are two textural varieties of staurolite in the sample: (a) Staurolite₁ occurs as a protected inclusion within kyanite (Fig. 2k). (b) Staurolite₂, often intergrown with quartz partially rims biotite or its aggregates enclose kyanite relics (Fig. 2n). In 09HNP-10, all the monazite grains occur in the foliation domain with shapes varying from elongated through ovoid to amoeboid.

4.2. Mineral chemistry

Mn-, Mg-, Ca- and Fe-X-ray element images of two representative garnet grains in the HNP-10-1 and HNP-10-2 show a uniform feature of Mn-rich core and Mn-poor rim and a rim-ward increase in Mg, Ca and Fe contents (Fig. 3a–h), which are also reflected in the compositional profiles (Fig. 3i–r). This rim-ward depletion in spessartine (12 → 07–08 mol%) and minor enrichments in pyrope (14 → 15) and X_{Mg} (16 → 17) (Table 1) contents in garnet are attributed to compositional zoning due to prograde garnet growth. One characteristic feature of garnet in this sample is the rim-ward enrichment in grossular content (02 → 04). The outermost garnet rim (~115–250 µm wide) in contact with biotite shows a fall in X_{Mg} (17 → 14) and grossular contents (04 → 03).

A garnet core in sample HNP-10-1 is enriched in heavy rare earth elements (HREE, Gd to Lu) relative to its inner rim (Table 2, Fig. 3s). Collectively and individually, all HREE elements and Y show a steady decrease in their concentrations from core to inner rim (Fig. 3t–v). The outermost garnet rim is, however, enriched in HREE and Y (Fig. 3t–u), which is also reflected in the changing Lu/Dy ratios from core/inner rim (~0.16–0.21) to outer rim (~0.88–0.93). We attribute the compositional zoning in garnet, reflecting core to rim depletions in HREE and Y contents, and broadly resembling a bell-shaped REE distribution, to prograde garnet growth (Hickmott and Shimizu, 1990; Otamendi et al., 2002). The spikes in HREE and Y at the outermost rim of garnet are explained by intragranular back-diffusion during garnet dissolution.

Biotite in the foliation domain is slightly more magnesian (0.57–0.59) than its occurrence as a partial rim around garnet (X_{Mg} = 0.56–0.57) (Table 1). Included biotite in garnet has X_{Mg} of 0.58–0.60. Both included plagioclase in garnet (An₁₁) and matrix plagioclase are oligoclase (An_{15–22}) with the matrix variety showing reverse zoning [An_{15–16}(C) → An_{21–22}(R)].

Matrix muscovite is generally sodic (Ms_{58–62} Pa_{26–29} Cel_{8–11} Prl_{02–04}) in composition. Paragonite has a composition of Pa₆₂ Cel₁₂ Ms₀₉ Prl₁₇.

Staurolite is compositionally homogeneous at X_{Mg} = 0.20. Matrix chlorite is more magnesian (X_{Mg} = 0.61) than that included within garnet (X_{Mg} = 0.50). Apatite inclusions in garnet are fluorapatite (X_F = 0.93) with minor hydroxylapatite content (X_{OH} = 0.06).

4.3. Mineral reaction history

We now use the textural and mineral compositional features to infer the possible reaction pathways that shaped the metamorphic evolution of the HGB. While both the samples record a common biotite + muscovite + staurolite + chlorite-defined foliation, mineral assemblage pre-

and post-dating the foliation is different and is differentially preserved in the two samples. In HNP-10, porphyroblastic garnet (together with its included phases such as biotite, chlorite and paragonite) (Fig. 2h), coarse staurolite aggregates (Fig. 2d) and muscovite with thick seams of fibrolitic sillimanite inclusions (Fig. 2e) have developed preceding the pervasive deformation event. Although of fragmentary in nature, these textural and compositional attributes when applied together attest to three distinct metamorphic stages: (a) a partially preserved prograde heating stage, as indicated by an early stability of chlorite (Fig. 2a, i), biotite (Fig. 2h) and paragonite (Fig. 2h), and formation of growth-zoned garnet (Fig. 3) following a model reaction, R1 (Table 3), (b) peak metamorphism, represented by garnet and staurolite, together with ilmenite, muscovite, biotite, sodic plagioclase and quartz, and (c) a garnet decomposition stage, producing biotite, sillimanite and relatively calcic plagioclase according to a model reaction, garnet + muscovite + plagioclase₁ → biotite + sillimanite + plagioclase₂ (R2a, Table 3). This reaction is supported by features such as partial biotite rim around garnet (Fig. 2i), development of fibrolitic sillimanite in muscovite (Fig. 2e), depletion in grossular contents in garnet rim and reverse zoning in plagioclase (An_{15–16} → An_{21–22}). Additional support for the decomposition of garnet comes from the trace element zoning in the mineral (Fig. 3s–v). In the garnet-absent domain, sillimanite can also be formed via the reaction, staurolite + muscovite + quartz → biotite + sillimanite + H₂O (R2b, Table 3). The occurrences of staurolite (Fig. 2c), chlorite (Fig. 2b) and sillimanite-free muscovite (Fig. 2j) in the foliation domain indicate a low-temperature hydration event, according to the model reaction, sillimanite + biotite + H₂O → chlorite + staurolite + muscovite (R3, Table 2).

Sample 09HNP-10, in contrast, has preferentially preserved a mineral reaction history following the foliation development. Textural features record a two stage mineral assemblage development in the rock: (a) the formation of kyanite (Fig. 2k) and rutile (Fig. 2l) from the decomposition of staurolite₁ (Fig. 2k) and ilmenite as part of a second kyanite-zone prograde burial and heating event and (b) retrograde regrowth of staurolite₂ (Fig. 2n) with or without chlorite after kyanite (Table 3 for model reactions). For ease of description, we name these two prograde burial events, pre- and post-dating the recorded deformation event as M₁ and M₂ respectively (Table 3). Summarising, the two investigated samples together record a polyphase metamorphic evolution of the terrane.

5. Calculation of metamorphic phase diagram and reconstruction of metamorphic P–T path

Even in the context of a possible polyphase metamorphic history of the HGB, this early stage of metamorphic evolution, which has mostly escaped from later equilibration is unlike that previously documented by Bouhallier (1995). We consider this ideally suited to explore and link with a possible older Mesozoic tectonothermal history of the terrane. For this, we utilize sample HNP-10 for its excellent record of multistage reaction history of the M₁ event, the focus of this study.

The phase diagram predicts an early stability of staurolite before garnet (at P < 7.5 kbar) and biotite during prograde heating along a medium-pressure-medium-temperature (MP-MT) thermal gradient, consistent with their thermal and baric stabilities in high-alumina pelite bulk rock composition as in HNP-10 (Fig. 4a). The P–T field of the M₁ peak metamorphic assemblage garnet + biotite + muscovite + staurolite + ilmenite + plagioclase + quartz is demarcated by the kyanite-in curve at higher T (T > 640–650 °C), the rutile-in curve at higher P (P > 8 kbar), the garnet-out and sillimanite-in curves at lower P (P between 6.0 and 6.8 kbar) and the paragonite-out curve at lower T (T < 600–640 °C) (Fig. 4b). Assuming the bulk rock has not been significantly fractionated prior to garnet growth, we constrain the P–T condition of the first appearance of garnet by knowing the compositions of the earliest stabilized prograde garnet core and applying the isopleth thermobarometry approach (after Vance and Mahar, 1998). Intersection of the three

Fig. 2. Back-scattered electron (BSE) images (a, h–i) and photomicrographs (b–g, j–n) of two representative pelitic schist samples (HNP-10 with thin sections numbering HNP-10-1 and HNP-10-2 and 09-HNP-10), showing key textural features. Mineral abbreviations used in this and other figures are after Kretz (1983). (a–j) Sample HNP-10 showing the development of a thinly banded schistosity (a–b) that wraps around Grt porphyroblast (c). Coarse aggregates of St (d) and locally Ms. (e) occur in the interfolial domains. The schistosity is overgrown by texturally late Bt (a), St (f) and Chl (g). Grt with rare inclusions of Bt, Pa (h), Qtz and Chl (i) is partially rimmed by Bt (a, g) at places. Whereas thick seams of fibrolite occur at the core of the interfolial Ms. (e), these are absent in the foliation-defining Ms. grains (j). Locations of monazite grains (M-1, M-2, M-3, M-4, M-5, M-6 and M-7) are marked in Fig. 2a. (k–n) Sample 09-HNP-10 showing the occurrences of large Ky porphyroblast, locally with included St (k) and Rt (l). The porphyroblasts overgrow a Bt + Ms. $\pm$ Chl-defined schistosity (m). Aggregates of a texturally later generation of St enclose relics of Ky (n).

Fig. 2 (continued).

Fig. 3. (a–h) X-ray element images in two representative Grt grains in thin sections HNP-10-1 (a–d) and HNP-10-2 (e–h). Also shown are the locations of selected compositional profiles in Grt grains [(i–m) and (n–r)]. (s) Chondrite normalised REE plots for different textural locations in garnet in sample HNP-10-1. (t–v) Core to rim compositional variations of trace elements Y, total HREE contents, Yb and Lu in garnet. Numbers refer to textural locations of the analytical spots in garnet, shown in Fig. (v). The Lu plot is magnified in Fig. (u).

compositional isopleths of garnet, X_{Prp} (Prp_{14}), X_{Grs} (Grs_{502}) and X_{Sps} (Sps_{12}) that are contoured in the calculated phase diagram, yields a P – T condition of ~6.4 kbar and 630 °C for the first appearance of garnet

in the rock (Fig. 4b). Due to relatively low modes of garnet (~1.5–3%) in the rock, we also consider that bulk rock fractionation due to garnet growth is unlikely to be significant (e.g. Marmo et al., 2002) making it

possible to use the inner rim composition of growth zoned garnet to estimate the peak P - T condition of metamorphism. The assumption, if valid, would yield tight intersections between the compositional isopleths of garnet and other phases (e.g. Evans, 2004).

As discussed in the mineral chemistry section, the Mg-Fe composition of the outer rim of growth-zoned garnet has been partially reset due to the decomposition of garnet (Fig. 3m,r). This implies that the maximum pyrope content in garnet outer rim is likely to be higher than that measured (Prp_{15.5}). We have estimated this pyrope content at ~16 mol% by projecting the core-rim pyrope profile up to a distance where X_{Grs} is at a maximum (Grs₀₄) (inset of Fig. 4b). We use these garnet (Prp₁₆ Grs₀₄ Sps₀₇) and matrix biotite ($X_{\text{Mg}} = 0.57$) compositions to calculate P - T conditions of peak metamorphism. Tight intersections of these compositional isopleths of garnet and biotite ($\Delta T \sim 15^\circ\text{C}$; $\Delta P \sim 0.3$ kbar) have yielded a P - T value of ~7.5 kbar, 640 °C (Fig. 4b), which is taken as the estimate of peak metamorphism in the HGB. The P - T estimate is in good agreement with the texturally constrained kyanite- and rutile-absent peak metamorphic assemblage.

We now apply the reconstructed metamorphic reaction pathways and estimated P - T conditions during garnet growth to deduce the metamorphic P - T path of evolution of the M_1 event. The dispositions of paragonite-out, chlorite-out and garnet-in curves and garnet mode contours reveal two segments of the prograde P - T path: (a) segment 1a, indicating heating accompanying loading intersects chlorite-out, paragonite-out and garnet-in curves in a sequence (Fig. 4c). This is supported by textural features such as the presence of chlorite and paragonite inclusions within garnet and complete absence of paragonite in the matrix. The phase diagram predicts a metamorphic sequence of an early stability of paragonite and chlorite and subsequently their complete removal in the mineral assemblages along path 1a (Fig. 4c). (b) Segment 1b with a steep dP/dT gradient marks a phase of prograde garnet growth, also supported by intersections of progressively higher garnet mode contours (Fig. 4c). There are three possible P - T pathways of evolution following the growth of garnet (Fig. 4c): (a) a phase of near-isobaric cooling (Path 2), (b) a segment of heating accompanying decompression followed by decompression and cooling (Path 3) and (c) a segment of near-isothermal decompression (Path 4). Path 2, which intersects increasing garnet mode contours can be discounted as it does not explain the post-peak reaction history of decomposition of garnet (Fig. 4c) and appearance of sillimanite (Fig. 2e). Path 3 is also unlikely as the segment of heating at lower pressure will involve the formation of kyanite, which is not recorded in the sample. We favour option (c) since the steep decompression path (path 4) intersects progressively decreasing garnet and muscovite and increasing biotite and plagioclase mode contours (Fig. 4c), culminating in the appearance of sillimanite and not kyanite at the lowest pressure end. This is consistent with the construed post-peak reactions (R2a,b, Table 3). Reappearance of chlorite and staurolite via reaction R3 (Table 3) in the foliation domain reflects a low- T hydration stage, which can be part of the

same retrograde P - T path and marking a phase of post-decompressional cooling. Alternatively, the syntectonic formation of chlorite and staurolite can be related with the temporally unrelated M_2 event. Nevertheless, our study reveals that this progressive change in P - T vectors from heating accompanying loading (Path 1a) through isothermal burial (Path 1b) and finally to isothermal decompression (Path 4) defines a clockwise P - T path of evolution of the M_1 event in the HGB (Fig. 4c).

6. Monazite EPMA dating

6.1. Texture, compositional zoning and geochronological results

We have analysed a total of 97 spots in 16 monazite grains from 3 thin sections [7 grains (M1 to M7) in HNP-10-1, 4 grains (M1 to M4) in HNP-10-2 and 5 grains (M1 and M3 to M6) in 09HNP-10] belonging to 2 samples of pelitic schist (HNP-10 and 09-HNP-10). The analyses reveal monazite grains of two age groups: (a) Mesoarchaeon monazite grains generally occur as inclusions in garnet (e.g. Fig. 5a for a representative monazite grain M1; see Supplementary Fig. S1 for other examples) and rarely in the matrix (e.g. monazite grain M5, Fig. 5d). (b) Neoarchaeon to Earliest Palaeoproterozoic monazite grains characteristically occur as discrete elongated grains in the matrix, aligning parallel to the M_2 foliation domain (Fig. 5g, j). Cores of some of these monazite grains show the presence of remnant Mesoarchaeon monazite (Fig. 5h-i, k-l). We interpret the Neoarchaeon to Earliest Palaeoproterozoic monazite grains to have originated due to dissolution-reprecipitation of the Mesoarchaeon monazite and to have grown synchronously with the M_2 metamorphic event. In this study, we have presented the textural, compositional and age information of the Mesoarchaeon monazite grains (Figs. 5, 6, 7). The details of Neoarchaeon to Earliest Palaeoproterozoic monazite geochronology will be presented in a separate publication.

X-ray element (Y, Th and U) imaging (Supplementary Fig. S1 for details and Fig. 5b, e, h, k for representative Y-X-ray element images) and individual spot chemical analyses (Supplementary Table S1) together reveal compositional zoning in the Mesoarchaeon monazite grains. Despite compositional overlaps, textural and compositional features attest to two types of compositional domains, CD1 and CD2. Compositional variations in these domains are controlled by the operation of a dominant cheralite substitution vector ($\text{REE} = \text{Ca} + \text{Th}$) (Fig. 5m).

CD1 is documented only in HNP-10 in the cores of composite monazite crystals, generally occurring as inclusions in garnet (Fig. 5a), matrix quartz from the interfolial domain (Fig. 5d) and staurolite from the foliation domain (Fig. 5g). The domain is marked by low-moderate Th ($\text{ThO}_2 = 2.5\text{--}6.0$ wt%), low-moderate Y ($\text{Y}_2\text{O}_3 = 1.2\text{--}1.5$ wt%) (Fig. 5b, e, h, k), very low U ($\text{UO}_2 = 0.2\text{--}0.4$ wt%) contents (Fig. 5n and Supplementary Fig. S1 for complete information), and coupled relatively higher Th/U ratios (8–19) but lower X_{YPO_4} contents (Fig. 5o).

Table 3

A summary of key metamorphic mineral assemblages, mineral compositions and mineral reaction history in metapelitic schists from the Hohenarsipur Greenstone Belt, WDC.

Sample/Met ^{lc} events/Bulk composition/location	Metamorphic assemblage	Mineral chemistry	Metamorphic reactions
HNP-10; M_1 - M_2 ; MnO = 0.08; Al ₂ O ₃ = 21.01; Ca [#] = 22; Mg [#] = 42; Lat:12°49'52", Lon:76°12'16"	M_{1P} : Pa, Bt, Qtz, Chl, (St,Ms); M_{1P} : Grt + St (+ Ms. + Bt + Pl + Ilm + Qtz); M_{1R} : Bt + Sil + Ca-Pl; M_2 (FOL): Chl, St, Bt, Ms., Qtz	Chl(FOL): 61; Chl (I ⁺ Grt):50; Chl(COR > Grt): 61; Bt(FOL): 57–59, 0.08–0.09; Bt(I ⁺ Grt): 58–60, 0.08–0.09; Bt(COR): 56–57, 0.08–0.12; Ms.(M): Pa ₂₆₋₂₉ Cel ₈₋₁₁ Ms ₅₈₋₆₃ Pr ₂₋₄ ; Pa(I ⁺ Grt): Pa ₆₂₋₆₈ Cel ₁₁₋₁₂ Ms ₀₉₋₁₅ Pr ₀₇₋₁₇ ; St(M): 19–20; Pl(I ⁺ Grt): An ₁₁ ; Pl (M): An ₁₅₋₁₆ (C) → An ₂₁₋₂₂ (R); Grt: Sps ₁₂ Grs ₀₂ Prp ₁₄ Alm ₇₂ (C) → Sps ₀₈ Grs ₀₄ Prp ₁₅ Alm ₇₄ (IR) → Sps ₀₈ Grs ₀₃ Prp ₁₃ Alm ₇₆ (E)	M_1 : Chl-out, Pa-in → Grt-in(R1) → Sil-in (R2a,b) M_2 : Chl-in (R3)
09-HNP-10; M_2 ; Lat:12°49'52", Lon:76°12'16"	M_2 (FOL): St, Chl, Ilm, Bt, Ms., Qtz; M_{2P} : Ky + Bt + Rt; M_{2R} : St + Ms		M_2 : Ky-in (R4) → St + Chl-in (R5)

Abbreviations used: Met^{lc}: Metamorphic; Lat: Latitude; Lon: Longitude; Al₂O₃ & MnO values in wt%; Ca[#] = 100*(Molar CaO/(Molar CaO + Molar Na₂O)); Mg[#] = 100*(Molar MgO/(Molar MgO + Molar FeO)); $M_{1/2P/R}$: Prograde/Peak/Retrograde metamorphic stages of M_1/M_2 events; FOL/M/COR: Foliation/Matrix/Corona; I: Inclusion; ^/>: Within/Against; C/R/E: Core/Rim/Edge; Mineral compositions expressed as, Values in Chl, St: 100*X_{Mg}; Values in Bt in sequence: 100*X_{Mg}, Ti in per formula unit per 11(O); R1 to R5: Metamorphic reactions 1 to 5, listed as follows: R1, St + Bt + Qtz → Grt + Ms. + V; R2a, Grt + Ms + Pl₁ → Bt + Sil + Pl₂; R2b, St + Ms. + Qtz → Bt + Sil + V; R3, Sil + Bt + V → St + Chl + Ms. ± Qtz; R4, St + Chl + Ms. ± Qtz → Ky + Bt + V; R5, Ky + Bt + V → St + Chl + Ms. ± Qtz.

Fig. 4. (a) MnNCKFMASHTi P-T pseudosection for a representative bulk rock composition of sample HNP-10, showing stability of equilibrium mineral assemblage fields and key index minerals. The P-T box indicates the possible P-T limits of mineral assemblage evolution in the investigated rock and shown in details in Fig. (b) and (c). (b) The mineral assemblage fields are contoured with compositional isopleths of X_{Spss} , X_{Grs} , X_{Prp} (for Grt core and inner rim compositions) and X_{Mg} (Bt). For the estimations of P-T conditions of prograde metamorphism, intersections of Grt compositional isopleths corresponding to growth zoned prograde Grt core are considered. The peak P-T condition is calculated by intersections of the isopleths of Grt inner rim composition (reconstructed X_{Prp} value shown in the inset) and X_{Mg} in matrix Bt away from its contact with Grt. (c) The same mineral assemblage fields in Fig. (b) are contoured with mode isopleths for Grt, Ms., Bt, Pl, St. The deduced metamorphic reaction pathways in HNP-10 are evaluated with modal variations of these minerals and possible metamorphic P-T paths of evolution (see text for details).

Spot monazite ages yield Mesoarchaeon dates with a range from 3246 to 3006 Ma (Supplementary Table S1; Fig. 5c, f, i).

CD2, which is recorded in both the samples occurs in several textural modes: (a) as thin rim around CD1 monazite (Fig. 5b–c, e–f), (b) as partial rim around CD1 monazite in a composite Mesoarchaeon-Neoarchaeon monazite grain (Fig. 5h–i), (c) as tiny relics within a Neoarchaeon monazite (Fig. 5k–l) and (d) as an oval-shaped core, being surrounded by an overgrowth of Neoarchaeon monazite. CD2 monazite as part of

composite CD1-CD2 grains generally occur as inclusions in garnet and in matrix quartz (Fig. 5a,d). In contrast, relics of CD2 core in composite Mesoarchaeon-Neoarchaeon monazite grains occur in the foliation domain. Compositionally, the domain contains moderately high Th ($\text{ThO}_2 = 1.4\text{--}5.0\text{ wt\%}$), higher Y ($\text{Y}_2\text{O}_3 = 1.4\text{--}1.9\text{ wt\%}$), moderate-high U ($\text{UO}_2 = 0.5\text{--}1.7\text{ wt\%}$) contents (Supplementary Fig. S1) and very low Th/U ratio (1–6) but relatively higher X_{YPO_4} contents (Fig. 5o). Spot monazite

Fig. 5. BSE and Y- X-ray element images and binary compositional plots of Mnz grains from pelitic schist samples HNP-10 and 09HNP-10. The analysed Mnz grains are compositionally and age zoned (Mesoarchaeoan and Neoarchaeoan to Earliest Palaeoproterozoic). In this study, we present the variations of Mesoarchaeoan compositional domains, here classified as CD1 (preserved in Mnz core) and CD2 (mostly as rims around CD1) (see text for further details). (a, d, g, k) Textural settings of analysed Mnz grains. (b, e, h, k) Y- X-ray element images showing differential preservations of CD1 and CD2 domains. (c, f, i, l) BSE images showing spot EMP ages and Th/U ratios. Spot colour codes for linked compositional domains used: Red, CD1 and Yellow, CD2. The uncertainties (2σ) of monazite spot ages are in the range of 63–169 Ma in HNP-10 and 100–179 Ma in 09-HNP-10. (m–o) Compositional plots for CD1 and CD2 compositional domains. (Th + U) vs. Ca(m), Th (ppm) vs. U (ppm) (n) and X_{YPO4} vs. Th/U (o) binary diagrams.

sample preparation. A bias from standard-sample matrix-mismatch (Fletcher et al., 2010; Hetherington et al., 2017; Stern and Berman, 2001) does not appear to be an issue here. Relevant instrumental indicators (UO^+/U^+ , secondary tuning parameters) were very similar between standards and unknowns. An analytical problem relating to sample preparation, i.e. the use of thin sections for the unknowns and epoxy casting for the standards is the most likely possibility, but even here there is no indication from instrumental conditions that this has caused a bias. Despite the issues with the detailed calibration of Pb/U and Pb/Th, the $^{207}Pb/^{206}Pb$ ratio is immune to these issues and also

provides better age resolution. The significance of these ages in a metamorphic context are further examined in the Discussion section.

8. Garnet-monazite thermometry

Textural and compositional features reveal that in a xenotime-absent mineral assemblage as in the HGB, monazite core (cf. CD1 compositional domain) grew in the presence of a growth-zoned garnet inner rim at metamorphic peak. These are: (1) This compositional type is preferentially preserved in monazite shielded either in the

Fig. 6. Probability density diagrams with histograms (a, c) and weighted mean ages of spot monazite ages from the CD1 and CD2 compositional domains of the investigated pelitic schist samples. All error bars are 2σ .

Fig. 7

Fig. 7. (a–d) X-ray Th images of Mnz grains (M1–5, 2, 7) showing the locations of in situ SHRIMP ^{207}Pb – ^{206}Pb ages (reported in Ma $\pm 1\sigma$) in HNP-10-1. (e) Probability density diagram showing two age populations in monazite, P1 at 3144 ± 7 Ma (error in 2σ) and P2 at 3105 ± 10 Ma. (f) Tera-Wasserburg concordia plot of monazite SHRIMP data. See text for details.

Table 4

SHRIMP monazite U–Pb isotope data in the pelitic schist sample HNP-10-1 from the Holenarsipur Greenstone Belt, WDC.

Label	Th/U	$^{204}\text{Pb}/^{206}\text{Pb}$ (e-5)	f_{206} (%)	$^{207}\text{Pb}/^{206}\text{Pb}$	$^{208}\text{Pb}/^{206}\text{Pb}$	$^{208}\text{Pb}/^{232}\text{Th}$	$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{235}\text{U}$	$^{238}\text{U}/^{206}\text{Pb}$	Age 8/32 (Ma)	Age 6/38 (Ma)	Age 7/6 (Ma)	%C (%)
M1-1	10.3 ± 2.5	1.94	0.03	0.2439 ± 15	2.584 ± 48	0.1867 ± 44	0.688 ± 26	23.12 ± 0.92	1.454 ± 56	3460 ± 74	3373 ± 102	3146 ± 10	107
M1-2	4.39 ± 0.49	0.71	0.01	0.2442 ± 19	1.081 ± 40	0.1736 ± 39	0.637 ± 29	21.44 ± 1.01	1.570 ± 71	3236 ± 67	3177 ± 115	3148 ± 13	101
M2-1	18.1 ± 3.2	1.13	0.02	0.2368 ± 15	4.531 ± 36	0.1710 ± 40	0.631 ± 30	20.61 ± 1.00	1.584 ± 75	3190 ± 69	3154 ± 119	3098 ± 10	102
M2-2	4.61 ± 0.34	1.20	0.02	0.2386 ± 10	1.125 ± 03	0.1711 ± 51	0.641 ± 25	21.08 ± 0.84	1.560 ± 61	3193 ± 89	3193 ± 100	3111 ± 07	103
M5-1	5.16 ± 0.26	0.23	0.00	0.2346 ± 22	1.288 ± 10	0.1733 ± 48	0.654 ± 26	21.15 ± 0.88	1.529 ± 60	3230 ± 83	3244 ± 101	3083 ± 15	105
M5-2	9.26 ± 0.33	0.56	0.01	0.2430 ± 07	2.264 ± 22	0.1701 ± 47	0.650 ± 25	21.78 ± 0.88	1.539 ± 61	3174 ± 82	3228 ± 102	3140 ± 05	103
M5-3	10.7 ± 1.6	0.12	0.00	0.2452 ± 13	2.628 ± 23	0.1773 ± 43	0.671 ± 33	22.69 ± 1.17	1.490 ± 75	3299 ± 74	3311 ± 132	3154 ± 09	105
M7-1	13.14 ± 0.56	2.94	0.05	0.2392 ± 30	3.226 ± 24	0.1688 ± 50	0.639 ± 23	21.09 ± 0.82	1.564 ± 55	3153 ± 87	3186 ± 90	3115 ± 20	102

Abbreviations: f_{206} – fraction of common ^{206}Pb in total Pb; %C – Concordance (%).

inner rim of garnet or in low-strain interfolial domain in the matrix. (2) Chondrite normalised REE plots show steep LREE enrichment and HREE depletion and a strong negative Eu anomaly (Fig. 8a), consistent with their growth in the presence of plagioclase as indicated by the reconstructed peak metamorphic assemblage. (3) Monazite cores showing relatively lower Y concentrations (Fig. 5o) and REE partitioning relationship between monazite and inner garnet rim, broadly in agreement with natural rock data (Fig. 8b) together reveal broadly co-eval crystallization of garnet and monazite (e.g. Hermann and Rubatto, 2003; Warren et al., 2018). We combine these compositional features of monazite and garnet with the composition of included apatite ($X_{\text{OH}} = 0.06$) in garnet and relatively sodic plagioclase composition ($X_{\text{An}} = 0.11$ to 0.15) either from its occurrence as inclusion within garnet or from matrix plagioclase core. The peak P–T conditions at which the fugacity of H_2O was calculated were acquired from the results of pseudosection modelling. The selected mineral compositions [X_{YAG} , X_{Grs} , X_{YPO_4} (Mnz), X_{OH} (Ap) and X_{An}] and calculated $f_{\text{H}_2\text{O}}$ value are listed in Table 5. Garnet–monazite thermometry has yielded temperatures of 605–610 °C (Table 5), in broad agreement with the estimated peak metamorphic temperature at ~640 °C.

9. Discussion

9.1. Timing of monazite growth and recrystallization behavior in a metamorphic context

Our study indicates formation of two distinct types of Mesoarchaean monazite grains. Textural and compositional features and garnet–monazite thermometric results together suggest that the CD1 monazite core formed during peak M_1 (cf. M_{1P}) metamorphism (Fig. 9). Textural and compositional features of the occurrence of CD2 monazite rims around the core, outer boundary of the former often resembling the outline of the remnant core (e.g. monazite grains M1 and M5, Fig. 5a–f) and

relatively sharp compositional variation (e.g. in Y, Th and U) across the domain boundary attest to the formation of the CD2 monazite via a mechanism of dissolution–recipitation of the CD1 monazite (cf. Bhowmik et al., 2014; Putnis, 2002). Their relatively higher Y contents and higher uranium and low Th/U compositions relative to their core (Fig. 5n–o) when combined with Y and HREE enrichments in the outer rim of garnet (Fig. 3s–v) indicate that this metamorphic recrystallization event was associated with the coupled decompositions of garnet and a uranium-bearing phase, most likely zircon. On the basis of this evidence, we relate the CD2 monazite rim with the post-peak decompression (M_{1R} metamorphic stage, where R refers to retrogression, Fig. 9).

Monazite Th–U–Pb EPMA ages have constrained the timings of development of monazite cores and rims at 3120 ± 28 Ma and 3094 ± 24 Ma respectively. Although, these ages are statistically the same within their error limits, they do provide a first order constraint on the timing of a Mesoarchaean metamorphic event between 3.12 and 3.09 Ga in the Holenarsipur Greenstone Belt. The higher precision of the SHRIMP ^{207}Pb – ^{206}Pb monazite dating further resolves these events at 3144 ± 7 Ma and 3105 ± 10 Ma. The older age domain is largely derived from monazite cores, which have been shielded from later equilibrations in garnet inner rim and interfolial matrix quartz. For this reason, we consider it as the timing of the M_{1P} metamorphism. For the younger monazite age domain, two of the analysed SHRIMP spots lie on monazite included in the outermost garnet rim. This part of the growth zoned garnet rim is compositionally modified due to its dissolution following peak metamorphism. The remaining two analyses of monazite are from the matrix. There are two possibilities to interpret the 3.11 Ga monazite age domain: (a) a reset M_{1P} age due to modifications of isotopic compositions during the retrograde metamorphism, (b) a variation in M_{1P} age due to an initial compositional heterogeneity of the monazite core domain. Since there is a difference of ~40 Myrs in the two age populations, which is larger than the age uncertainty, we discount option

Fig. 8. (a) Chondrite normalised REE plot of Mnz core from different textural locations in the pelitic schist sample, HNP-10-1. (b) REE element partitioning relationship between matrix monazite core and garnet inner rim (dark blue band) is compared with published data (cyan band). Abbreviations used: HR03, Hermann and Rubatto (2003); B06, Buick et al. (2006); R06, Rubatto et al. (2006); W18, Warren et al. (2018).

Table 5
Results of garnet-monazite thermometry in sample HNP-10-1.

Input parameters	Values
X_{Grs}	0.04
X_{YAG}	0.00015
X_{An}	0.11–0.15
X_{YPO4}	0.03
$X_{\text{OH,Ap}}$	0.06
$f(\text{H}_2\text{O})$	6623.6
$\text{Ln}(K_{\text{Eq}})$	8.278–8.665
$T_{\text{Calc}} (^{\circ}\text{C})$	605–610
Textural settings	Grt (IR), Mnz (C), Ap(I ⁺ Grt), Pl (I ⁺ Grt, M(C))

Textural abbreviations as in Table 1; Mineral mole fraction data computed after Pyle et al. (2001).

(b). Based on this analysis, we consider the best estimate of the peak M_1 metamorphism in the Holenarsipur Greenstone Belt is 3.14 Ga with the subsequent decompression recorded at 3.11 Ga.

In a previous study, Bouhallier (1995) established a Neoarchean to Earliest Palaeoproterozoic metamorphism in the HGB on the basis of garnet Sm–Nd ages (2473 ± 10 Ma to 2527 ± 34 Ma) in two pelitic schist samples (quoted in Jayananda et al., 2013), which are also recorded in our investigated pelitic schist samples. Published garnet Sm–Nd (Bouhallier, 1995; Li et al., 2018) and U–Pb zircon (Li et al., 2018) ages in the high-grade Sargur supracrustals and U–Pb zircon ages in granulite facies crustal blocks adjoining the Dharwar Craton (e.g. Brandt et al., 2014) also indicate that vast stretches of the Western Dharwar Craton and also the adjoining Southern Granulite Terrane have recorded a pervasive tectonothermal event in Neoarchean to Earliest Palaeoproterozoic time that reached high-pressure granulite facies in the south. Our study reveals that despite a veil of a strong amphibolite facies kyanite-zone metamorphic overprint, an earlier Mesoarchean metamorphism between 3.14 and 3.11 Ga can still be recognised in localised domains of the HGB. While several previous studies inferred Mesoarchean metamorphism in the WDC (Jayananda et al., 2013, 2015; Raith et al., 1982), this is the first comprehensive documentation of the event in the terrane, including reconstruction of the metamorphic P–T path.

9.2. Mesoarchean metamorphic P–T Path and geodynamic significance

The reconstructed CW metamorphic P–T path in the HGB reveals a two-stage prograde evolution (M_{1PR} metamorphic stage): an initial phase of heating accompanying burial along a relatively gentle dP/dT gradient (~ 30 bar/ $^{\circ}\text{C}$) followed by burial ($\Delta P \sim 1.1$ kbar) along a steep dP/dT gradient (~ 110 bar/ $^{\circ}\text{C}$), culminating in peak metamorphism (M_{1P}) at 7.5 kbar, 640 $^{\circ}\text{C}$ (Fig. 9). In a general way, such a change in dP/dT gradient reflects a change in burial mechanism from one in which the underthrust or buried Holenarsipur Greenstone Crust equilibrated with relatively warmer footwall and hanging wall rocks to one of extremely rapid burial. The retrograde isothermal decompression path (M_{1R} stage) which broadly retraces the steep prograde burial path, but at higher temperatures also implies a very short-lived residence at metamorphic peak and a very fast tectonically-driven exhumation. Although such a hair-pin P–T path, reflecting a very rapid burial and exhumation cycle, is indicative of a subduction zone metamorphic setting, particularly in Phanerozoic metamorphic belts (e.g. Agard et al., 2009; Ao and Bhowmik, 2014; Ernst, 1988), the application of the same in Archaean granite-greenstone terrane is controversial (e.g. see Sizova et al., 2018 for a review).

In the Palaeo- to Mesoarchean Barberton Granite-Greenstone Terrane (BGGT), several studies have recorded high-pressure (P_{Max} from ~ 8.5 to 12 kbar) amphibolite facies metamorphism along clockwise metamorphic P–T paths at ca. 3200 Ma (Fig. 9) and an oblique crustal section in the Slotzberg terrane (Cutts et al., 2014; Diener et al., 2005; Dziggel et al., 2002; Moyen et al., 2006). This has been explained by a model of crustal thickening due to an accretionary tectonic process

(Cutts et al., 2014 and references cited therein). Several recent studies, involving thermo-mechanical modelling (e.g. Sizova et al., 2015, 2018) with or without integrated metamorphic and geochronological constraints of natural rock data (e.g. François et al., 2014), however, suggest that burial-exhumation cycles along clockwise P–T loops (e.g. Fig. 10) are not diagnostic of subduction-accretion tectonics. These can also be explained by vertical tectonics akin to sagduction settings (terminology Macgregor, 1951) (e.g. in ca. 3.5–3.2 Ga high-grade Mount Edgar high-grade metamorphic rocks in the East Pilbara Granite-Greenstone Terrane, EPGGT, François et al., 2014; BGGT, Van Kranendonk, 2011b, Van Kranendonk et al., 2014).

In this model, spatial association of Archaean granitoid domes and intervening volcano-sedimentary basins, now represented by greenstone terranes and metamorphic transformations in greenstone protoliths are explained by a co-eval process of gravitational sinking of dense greenstone cover sequences into narrow belts and emplacement of partially melted felsic crust into broad granitoid domes (François et al., 2014; Sizova et al., 2018). The model predicts a wide diversity of metamorphic P–T paths and peak P–T conditions of greenstone metamorphism depending on their spatial positions vis-à-vis the granitoid domes e.g. greenstone margin, greenstone interior and greenstone enclaves within the granitoids, (summarised in Fig. 10a–c). The model also calculates a large range of apparent thermal gradients, including significantly lower gradients (ca. 15 $^{\circ}\text{C}/\text{km}$) (François et al., 2014), akin to those proposed for modern and Archaean subduction-accretion zones (e.g. Moyen et al., 2006). Thus in the EPGGT, adjoining Mount Edgar dome, greenstone enclaves in the dome interior (sample PB-11-23, Fig. 9) were extensively heated at shallow to mid-crustal levels to produce metatexite migmatite (see also Fig. 10c for thermo-mechanical model prediction). In contrast, rocks in the greenstone interior (e.g. samples PB-11-35 and 33, Fig. 9) were progressively cooler away from their contact with the granitoid dome (Fig. 10b). Some of the rocks from the margin were also buried to a deeper level (sample No. PB-11-23, Figs. 9, 10a). While these thermo-mechanical models with initial conditions appropriate for vertical tectonics in Eoarchean to Mesoarchean time period, reproduce geometry of metamorphic P–T paths (e.g. Sizova et al., 2018) that have been obtained from natural rock data (e.g. Fig. 9), one key distinction of sagduction from subduction is that the former is a mechanically short-lived process. The burial and exhumation cycle of the greenstone crust in sagduction is broadly

Fig. 9. P–T diagram showing a comparison between the P–T path from this study and published Mesoarchean P–T paths and peak metamorphic conditions from two well studied greenstone belts, namely the Barberton Greenstone Belt (BGB) and East Pilbara Granite-Greenstone Terrane (EPT).

synchronous with peak magmatic activities (François et al., 2014). This is followed by protracted slow thermal relaxation periods.

9.3. Mesoarchaean tectonics in the Western Dharwar Craton

The pelitic schist lies at the interface between the Holenarsipur Greenstone Belt and the Halekote trondhjemite pluton (Fig. 1b). The latter was emplaced into the supracrustal sequence as a diapiric body (Bouhallier et al., 1995). While Bouhallier et al. (1995) applied field observations and structural criteria to propose a sagduction model for the evolution of the Holenarsipur Greenstone Belt, which has been widely accepted since then as a general geodynamic model in the Mesoarchaean, it was never independently validated by co-eval metamorphic age constraints from the sagducted greenstones. The metamorphic and chronological constraints in this study help to resolve this issue.

A collation of published zircon and titanite U-Pb geochronological data of felsic magmatic bodies reveal three periods of magmatism in the terrane (Fig. 11): (1) between 3411 and 3276 Ma (Beckinsale et al., 1980, 1982; Bhaskar Rao et al., 2008; Guitreau et al., 2017; Jayananda et al., 2015; Peucat et al., 1993; Santosh and Li, 2018) marking the main phase of growth of the Garur and Eastern TTG gneisses, (2) between 3230 and 3200 Ma (Jayananda et al., 2015) with a precise time constraint of ~3178 Ma as the timing of emplacement of Halekote and equivalent trondhjemite pluton (Guitreau et al., 2017) and (3) between 3134 and 3103 Ma as a minor phase of granite intrusion in the Garur gneiss (Guitreau et al., 2017; Jayananda et al., 2015). Published U-Pb detrital and inherited zircon

ages and Nd-Hf-Pb isotope data additionally indicate a cryptic record of an even older phase of crust formation in the WDC dating back to ~3.8 Ga (Bhaskar Rao et al., 2008; Guitreau et al., 2017; Hokada et al., 2013; Jayananda et al., 2015; Lancaster et al., 2015; Nutman et al., 1992; Sarma et al., 2012).

On the basis of Hf isotopic fingerprinting of detrital zircons, Guitreau et al. (2017) related these crust-formation events in the WDC with specific crust/mantle reservoirs: (a) Pre-3400 Ma crust sourced from several distinct mantle domains with ancient depletion history. (b) 3400–3200 Ma felsic crust originating from a common mantle source of mildly-depleted to near-chondritic Hf isotopic compositions with limited or no continental influence. (c) Magmatic events from ~3200 Ma onwards with signatures of reworked oldest WDC crust and also of dominant lithologies. This shift in isotopic reservoirs from predominantly mantle to continental types when evaluated with theoretical results, thermo-mechanical model predictions (François et al., 2014; Sizova et al., 2015, 2018) and natural rock data (e.g. Jayananda et al., 2008, 2015; Guitreau et al., 2017 and this study) suggests a change in the style of Mesoarchaean tectonics in the WDC from ca. 3.2 Ga onwards.

While there are suggestions that subduction tectonics may have led to the growth of TTG crust in the Gorur-Holenarsipur region (e.g. Jayananda et al., 2015; Naqvi et al., 2009), the overall geological setting and geochemical framework of the terrane [e.g. Sargur komatiite eruption as part of a plume-driven oceanic plateau formation event (e.g. Jayananda et al., 2008)] and spatial association of Holenarsipur Greenstone keel and Halekote diapirs (e.g. Bouhallier et al., 1995) are

Fig. 10. Metamorphic P-T-t paths from thermo-mechanical model predictions (after Sizova et al., 2018) for two end member thermo-tectonic settings: sagduction (a-c) and reworking of partially exhumed sagducted greenstone (d). (a) Greenstone margin against dome. (b) Greenstone interior. (c) Greenstone enclaves in granitoids. For a comparison, the P-T path of evolution of the Holenarsipur Greenstone Belt has also been shown.

Fig. 11. A summary diagram showing a comparison of Mesoarchaeoan magmatic and metamorphic events in different litho-tectonic domains of the WDC. The numbers in the figure represent references: 1, Jayananda et al. (2015); 3, This study; 4, Peucat et al. (1995); 5, Jayananda et al. (2008), 6, Nutman et al. (1992).

supportive of a model of gravity-driven vertical tectonics in a time frame between 3.4 and 3.2 Ga. Seen in that context, despite the remarkable similarity of the Holenarsipur Greenstone P-T path with that from sagduction model simulation (Fig. 10a), the estimated peak greenstone metamorphism (this study) is younger in time than the diapiric plutonism by ≥ 35 Myrs, but broadly coincident with metamorphism in associated TTG gneisses and Halekote diapir (Fig. 11). The metamorphism locally has resulted in crustal anatexis and granite vein formation (Jayananda et al., 2015). This raises the possibility that the tectonics responsible for the rapid burial and exhumation of a segment of the Holenarsipur Greenstone also reworked host TTG gneisses and diapirs. As stated above, thermo-mechanical models also predict formation of reworked (accreted) crust of strongly deformed and poly-phase metamorphosed (Fig. 10d) granite–greenstone and subduction-related sequences at later stages of sagduction tectonics as it evolves from predominantly vertical to horizontal shortening modes (e.g. Sizova et al., 2015, 2018). While the inferred model and reconstructed P-T paths and peak P-T estimates differ, in part due to chosen model physical parameters, this inference of a tectonic transition is also consistent with zircon Hf isotopic studies that reveal contributions of reworked older crust in the generations of post 3.2 Ga felsic crust in the WDC (Guitreau et al., 2017). On the basis of these arguments, we propose that this first comprehensive record of ca. 3.14–3.11 Ga Mesoarchaeoan metamorphism in the HGB is part of a pervasive tectono-thermal event during late stages of sagduction tectonics. The event has led to large-scale cratonization of the WDC. Future studies should be directed

to constrain the metamorphic and tectonic framework of this Mesoarchaeoan tectono-thermal event from different litho-tectonic elements of the Western Dharwar Craton.

10. Conclusions

1. Rare high-alumina pelitic schists from the ca. 3.3–3.1 Ga HGB in the WDC, South India record a near complete preservation of an older Mesoarchaeoan metamorphic history, despite a strong superposed metamorphic overprint in the latest Neoproterozoic to earliest Palaeoproterozoic time period.
2. The older history, now protected in low-strain domains of these rocks reveals three distinct metamorphic stages, which in a sequence are: (a) a partially preserved prograde heating stage, which is indicated by an early stability of chlorite ($X_{Mg} = 0.50$), biotite ($X_{Mg} = 0.58$ – 0.60) and paragonite ($Pa_{62-68}Cel_{11-12}Ms_{09-15}Pr_{107-17}$) and formation of growth-zoned garnet [$Sp_{512}Gr_{502}Pr_{14}Alm_{72}$ (Core) $\rightarrow$ $Sp_{507}Gr_{504}Pr_{16}Alm_{73}$ (Rim)], (b) the peak metamorphism, which is represented by an assemblage of garnet + staurolite ($X_{Mg} = 0.20$) + ilmenite + muscovite ($Pa_{26-29}Cel_{8-11}Ms_{58-63}Pr_{12-4}$) + biotite ($X_{Mg} = 0.57$ – 0.59) + sodic plagioclase (An_{11} to An_{15-16}) + quartz, and (c) the retrograde metamorphism, which is marked by the decomposition of garnet and formation of biotite ($X_{Mg} = 0.56$ – 0.57), fibrolitic sillimanite and calcic plagioclase (An_{21-22}).
3. On the basis of P-T pseudosection modelling in the MnTiNCKFMASH chemical system, we interpret the three metamorphic stages in terms of a hairpin clockwise P-T trajectory. The trajectory shows a two-stage prograde segment with an initial phase of heating accompanying burial (dP/dT gradient ~ 30 bar/ $^{\circ}C$) and a later phase of steep burial ($\Delta P \sim 1.1$ kbar, dP/dT gradient ~ 110 bar/ $^{\circ}C$), peak metamorphism at 7.5 kbar, 640 $^{\circ}C$ and a post-peak, steep decompression stage ($P \sim 6.3$ kbar at 635 $^{\circ}C$).
4. Chemically zoned monazite grains shielded within garnet or occurring in low-strain matrix domains attest to two stages of crystallization history during the Mesoarchaeoan metamorphism: (a) Monazite cores with relatively lower Y concentrations, higher Th/U ratios and a strong negative Eu anomaly crystallized in the presence of a growth-zoned garnet inner rim at metamorphic peak. This interpretation is additionally supported by the application of garnet-monazite trace element thermometry. The combinations of monazite core ($X_{YPO4} = 0.03$), garnet inner rim ($X_{GrS} = 0.04$; $X_{YAG} = 0.00015$), included apatite ($X_{OH} = 0.06$) in garnet and relatively sodic plagioclase composition ($X_{An} = 0.11$ to 0.15) have yielded temperatures of 605–610 $^{\circ}C$, in broad agreement with the estimated peak metamorphic temperature at ~ 640 $^{\circ}C$. (b) Monazite rims with higher Y and uranium compositions and lower Th/U ratios formed during the post-peak decompression stage associated with coupled decompositions of garnet and zircon.
5. The EPMA monazite dating has yielded mean core ages at 3120 ± 28 Ma and rim ages at 3094 ± 24 Ma. Although, these ages are statistically the same within their error limits, they do provide a first order constraint on the timing of a Mesoarchaeoan metamorphic event between 3.12 and 3.09 Ga in the HGB.
6. The SHRIMP ^{207}Pb – ^{206}Pb monazite dating method has yielded monazite core and rim ages at 3144 ± 7 Ma and 3105 ± 10 Ma respectively. The higher precision of the SHRIMP monazite ages allows resolution of the timings of peak and retrograde metamorphism in the terrane at 3.14 and 3.11 Ga respectively.
7. Our findings, when evaluated with the available geological, isotopic, geochronological data of Meso- to Palaeoproterozoic felsic crust as well as thermo-mechanical models suggest that the Mesoarchaeoan metamorphism is part of a widespread tectono-thermal event in the Western Dharwar Craton. The event marks a shift in tectonic style from initial sagduction to later horizontal plate motions.

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