New LA-ICPMS U–Pb ages of detrital zircons from the Highland Complex: insights into late Cryogenian to early Cambrian (ca. 665–535 Ma) linkage between Sri Lanka and India

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ABSTRACT
Here we report new LA-ICPMS U–Pb zircon geochronology of ultrahigh temperature (UHT) metasedimentary rocks and associated crystallized melt patches, from the central Highland Complex (HC), Sri Lanka. The detrital zircon $^{206}\text{Pb}/^{238}\text{U}$ age spectra range between 2834 ± 12 and 722 ± 14 Ma, evidencing new and younger depositional ages of sedimentary protoliths than those known so far in the HC. The overgrowth domains of zircons in these UHT granulites yield weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age clusters from 665.5 ± 5.9 to 534 ± 10 Ma, identified as new metamorphic ages of the metasediments in the HC. The zircon ages of crystallized in situ melt patches associated with UHT granulites yield tight clusters of weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages from 558 ± 1.6 to 534 ± 2.4 Ma. Thus, using our results coupled with recently published geochronological data, we suggest a new geochronological framework for the evolutionary history of the metasedimentary package of the HC. The Neoarchean to Neoproterozoic ages of detrital zircons indicate that the metasedimentary package of the HC has derived from ancient multiple age provenances and deposited during the Neoproterozoic Era. Hence, previously reported upper intercept ages of ca. 2000–1800 Ma from metaigneous rocks should be considered as geochronological evidence for existence of a Palaeoproterozoic igneous basement which possibly served as a platform for the deposition of younger supracrustal rocks, rather than timing of magmatic intrusions into the already deposited ancient sediments, as has been conventionally interpreted. The intense reworking of entire Palaeoproterozoic basement rocks in the Gondwana Supercontinent assembly may have caused sediments of multiple ages and provenances to incorporate within supra-crustal sequences of the HC. Further, our data supports a convincing geochronological correlation between the HC of Sri Lanka and the Trivandrum Block of Southern India, disclosing the Gondwanian linkage between the HC of Sri Lanka and Southern Granulite Terrain of India.

ARTICLE HISTORY
Received 30 December 2015
Accepted 30 April 2016

KEYWORDS
LA-ICPMS U–Pb geochronology; Gondwana; Highland Complex; Sri Lanka; Southern Granulite Terrain

Introduction

Sri Lanka represents an integral part in the centre of east Gondwana and hence is an important component of the record of collision and amalgamation of Gondwana. However, the precise geotectonic position of Sri Lanka in East Gondwana with respect to other supercontinent fragments is still controversial (e.g. Kröner et al. 2003; Kehelpannala 2004; references therein). Therefore, it is essential to expand the high-precision geochronology on different types of rocks to explore the capability of correlation of Sri Lanka with its neighbouring Gondwana terrains such as southern India and east Antarctica. The recent studies on metaigneous rocks from Sri Lanka have revealed prominent Neoproterozoic tectonomagmatic events (e.g. Santosh et al. 2012, 2014; He et al. 2015), which closely correlate the basement of the island with neighbouring Gondwana terrains such as Southern Granulite Terrain of India (e.g. Collins et al. 2007b; Teale et al. 2011; Kröner et al. 2012; Plavsa et al. 2014) and the Lutzów–Hölhm Complex, Antarctica (e.g. Satish-Kumar et al. 2008; Tsunogae et al. 2014, 2015; Takamura et al. 2015).

In this study we present LA-ICPMS U–Pb zircon geochronology data of four ultrahigh temperature metasedimentary granulites and two in situ melt...
patches associated with the ultrahigh temperature (UHT) granulites, collected from the central Highland Complex, Sri Lanka (Figure 1). Using the results from this study and available age data, we construct a new geochronological framework for the evolution of the metasedimentary package of the Highland Complex. Further, we provide new insights into possible Gondwana linkage of the Highland Complex of Sri Lanka to the southern granulite terrain of India.

**Geological background**

**Overview of general geology and petrology of Sri Lankan basement**

Approximately 90% of Sri Lanka is underlain by granulite to upper amphibolites facies rocks with minor Miocene limestone and minor Jurassic and recent deposits (Cooray 1994). Early workers subdivided the Sri Lankan basement rocks into a number of units (e.g. Adams 1929; Coates 1935; Vitanage 1972; Cooray 1984, 1994; Kröner et al. 1991); however, the currently accepted classification relies largely on Nd-model ages and zircon U–Pb dating (Milisenda et al. 1988, 1994; Liew et al. 1991) as presented in Cooray (1994), namely, from the west to the east the Wanni Complex (WC), the Kadugannawa Complex (KC), the Highland Complex (HC), and the Vijayan Complex (VC) (Figure 1(a)).

**The Highland Complex**

The HC contains granulite facies metasedimentary and metagneous rocks including quartzites, marbles, calc-silicates, pelitic gneisses, charnockites, and orthogneisses (Cooray 1994; Mathavan and Fernando 2001). However, in the southwestern part of the HC the metasediments such as marble and quartzite are scarce. Instead, thin bands of wollastonite-, scapolite- and diopside-bearing calc-silicates and cordierite-bearing gneisses are dominant (Hapuarachchi 1968; Cooray 1984; Perera 1984; Prame 1991; Mathavan et al. 1999). Earlier workers defined the HC as a tilted crustal section with a metamorphic pressure and temperature ($P$–$T$) gradient across the Complex, from roughly 4.5–6 kbar and 700–750°C in the southwest to 8–9 kbar and 800–900°C in the east and southeast (Faulhaber and Raith 1991; Raase and Schenk 1994; Schumacher and Faulhaber 1994; Kriegsman 1996; Mathavan et al. 1999; Malaviarachchi and Takasu 2011a; Dharmapriya et al. 2014b). Sajeev and Osanai (2005) defined a thermal gradient of ~600–1000°C across southwestern to the central part of the HC.

During the last two and a half decades, evidence for UHT metamorphism has been reported from a few localities in the central HC. The $P$–$T$ estimates for these UHT granulites revealed $T$ of 925–1150°C at pressures of 9–12.5 kbar (e.g. Osanai 1989; Osanai et al. 2000; 2006; Kriegsman and Schumacher 1999;
Bolder-Schrijver et al. 2000; Sajeev and Osanai 2004a; 2004b; Sajeev et al. 2007; Malaviarachchi and Dharmapriya 2015; Dharmapriya et al. 2015a; b).

The Wanni Complex
This Complex is comprised of upper amphibolite to granulite facies metagneous rocks with minor metasediments. The metagneous rocks range from granitic, granodioritic, monzonitic, tonalitic, charnockitic, and enderbitic composition (Pohl and Emmermann 1991). The minor metasedimentary rocks are composed of garnet-sillimanite gneisses, cordierite gneisses, quartzites, and calc-silicate rocks, which occur close to the inferred tectonic boundary with the HC (Kehelpannala 1997). The incipient charnockitization in the host amphibolite facies hornblende gneisses around Kurunegala are mainly interpreted as due to fluid influx along shear zone in the Sri Lankan lower crust (e.g. Hansen et al. 1987; Burton and O’Nions 1990; Baur et al. 1991). Estimated P–T conditions are\( P = 3.5–7.5 \text{kbar} \) and \( T = 600–900^\circ\text{C} \) (Faulhaber and Raith 1991; Schenk et al. 1991; Weerakoon et al. 2001). The WC contains unmetamorphosed post-tectonic granite at Tonigala (e.g. Cooray 1984, 1994; Hözl et al. 1991) and carbonatite deposits at Eppawala and Kavisigamuwa (e.g. Pitawala and Lottermoser 2012; Madugalla et al. 2014). Migmatization is also widespread through the Complex (e.g. Kriegsman 1993).

The Vijayan Complex
The VC is predominantly metamorphosed at upper amphibolite-facies, consisting of microcline-bearing granitic gneisses, augen-gneisses, migmatites, and amphibolite-facies, consisting of microcline-bearing granitic gneisses, augen-gneisses, migmatites, and hornblende-biotite gneisses (Cooray 1984; Kehelpannala 1997; Mathavan et al. 1999; Kröner et al. 2013). Rare sedimentary enclaves such as metaoquartzite, calc-silicate rocks, and marble are also found (Dahanayake 1982; Dahanayake and Jayasena 1983). The late-stage patchy charnockitization is affected in the area around Pottuvil (Jayawardena and Carswell 1976; De Maesschalck et al. 1990).

The Kadugannawa Complex
The KC, formerly called ‘Arenas’ (Vitanage 1972; Almond 1991), forms doubly plunging upright folds in the central part of the country. This unit mainly consists of ca. 900 Ma old hornblende- and biotite-bearing ortho-gneisses, gabbros, diorites, granodioritic to granitic gneisses, charnockites, enderbites, and minor metasediments (Kröner et al. 2003; Willbold et al. 2004). Kleinschrodt et al. (1991) described the KC as a basic layered intrusion at the deepest level. Kriegsman (1994) argued that these intrusions are locally overlain by a thin metasedimentary sequence and contains with calc-silicate, quartzite and itabirites. Based on geochronological, geochemical, and structural evidence some workers (e.g. Kehelpannala 1997; Kröner et al. 2003) suggested that the KC is a part of the WC. More recently, Santosh et al. (2014) argued that the KC is part of a disrupted huge arc magma chamber that was exhumed and transposed along the margin of the WC.

Brief summary of previous geochronological studies

The Highland Complex
The HC is the oldest litho-tectonic unit, with Nd-model age of 2000–3400 Ma (Milisenda et al. 1988, 1994). The first geochronological research on Sri Lanka was presented by Boltwood (1907) publishing the first \( U–Pb \) ages of ca. 2200 and 900 Ma in thorianites from Galle at the south coast of the southwestern part of the HC. The first detailed geochronological study of the HC was done by Crawford (1969) and Crawford and Oliver (1969) using \( Rb–Sr \) and \( K–Ar \) methods on whole-rock and mineral samples. The authors suggested that the supracrustal assemblage of the Highland Complex was of Archaean age and the high-grade metamorphism up to granulite facies took place ca. 2000 million years ago. Hözl et al. (1991, 1994) and Köhler et al. (1991) presented \( Rb–Sr \) whole rock and \( Sm–Nd \) garnet ages indicating ca. 2000 Ma. Kagami et al. (1990) reported the \( Rb–Sr \) and \( Sm–Nd \) isochron ages of 2600–2300 Ma and 500–450 Ma, respectively. Cordani and Cooray (1989) reported \( Rb–Sr \) isochron age of 1100 Ma and De Maesschalck et al. (1990) interpreted an \( Rb–Sr \) whole rock isochron age of 1930 Ma as the timing of high-grade metamorphism. Hözl et al. (1991) and Milisenda (1991) measured garnet-whole rock \( Sm–Nd \) ages from para-gneisses and metabasites in the HC and showed the metamorphic ages of ca. 600 Ma. Sajeev et al. (2003) reported a middle Proterozoic internal isochron age of ca. 1500 Ma using garnet core, whole rock, and felsic fraction (quartz + plagioclase) in \( Sm–Nd \) system, from a rock having a UHT assemblage. The ‘ordinary granulites’ of the HC show ca. 550 Ma metamorphic age.

The ages of the detrital zircon spectra are in the range of ca. 3200–2000 Ma (Kröner et al. 1987; Hözl et al. 1991, 1994) while metamorphic age has been interpreted as 610–550 Ma. Kröner et al. (1987) reported a Pb loss event around 1100 Ma from metasediment. Using \( U–Pb \) isotopes in ilmenite, Burton and O’Nions (1990) measured an isochron age of 1100 Ma, and reported as the metamorphic age. Kröner and Williams (1993) reported zircon \( U–Pb \) age of 1900 Ma for granitic gneiss from the Complex as the timing of crystallization of the igneous protolith and the lower intercept of
531 Ma as the age of metamorphism. Sajeev et al. (2007) reported U–Pb zircon metamorphic age of ca. 580 Ma from relatively HP/UHT mafic granulites. In the U–Pb system on zircon and monazite from the Sri Lankan UHT, metasedimentary rocks have ages clustering at 1700 Ma and at ca. 1400–830 Ma, while some of those having overgrowths of ca. 570 Ma (Sajeev et al. 2010). The CHIME dating of monazite by Malaviarachchi and Takasu (2011b) yielded a wide age range of ca. 728–460 Ma. Recently, geochronological studies on metagranites by Santosh et al. (2014), He et al. (2015) and Takamura et al. (2015) have reported late Neoproterozoic–Cambrian multiple thermal events in the HC. Further, Lu–Hf data of zircon of mafic and intermediate granulites and charnockites show Hf crustal model ages in the range of ca. 1500–2800 Ma (Santosh et al. 2014; He et al. 2015).

**The Wanni Complex**

The WC yields Nd-model ages of 1000–2000 Ma (Milisenda et al. 1988, 1994) while the intrusion ages of Wanni gneisses range from 790–750 Ma to 1100–1000 Ma. Kröner et al. (1994) reported a Pb–Pb evaporation age of 1330 Ma for detrital zircon indicating that the WC sediments were derived from Mesoproterozoic source rock. The unmetamorphosed granite at Tonigala and monazite from the Sri Lankan UHT, metasedimentary rocks have ages clustering at 1700 Ma and at ca. 1400–830 Ma, while some of those having overgrowths of ca. 570 Ma (Sajeev et al. 2010). The CHIME dating of monazite by Malaviarachchi and Takasu (2011b) yielded a wide age range of ca. 728–460 Ma. Recently, geochronological studies on metagranites by Santosh et al. (2014), He et al. (2015) and Takamura et al. (2015) have reported late Neoproterozoic–Cambrian multiple thermal events in the HC. Further, Lu–Hf data of zircon of mafic and intermediate granulites and charnockites show Hf crustal model ages in the range of ca. 1500–2800 Ma (Santosh et al. 2014; He et al. 2015).

The Kadugannawa Complex

The KC yields Nd-model ages ranging between 1000 and 2000 Ma (Milisenda et al. 1988, 1994). Kröner et al. (1987) reported SHRIMP zircon U–Pb age of ca. 1100 Ma. Pb–Pb zircon evaporation ages showed ca. 770–1100 Ma indicating multiple calc-alkaline magmatic activities in the KC (Kröner et al. 2003; Willbold et al. 2004). Most of the dioritic to granodioritic rocks yielded ca. 900 Ma ages. Perera and Kagami (2011) argued that the regional high-grade metamorphism of the KC must be older than 1100 Ma based on a study of Sr–Nd isotope systematics on charnockitic rocks. Santosh et al. (2014) and He et al. (2015) reported U–Pb zircon ages yielding 980–920 Ma as early Neoproterozoic magmatism and metamorphism at 530 Ma. Zircons in their garnet-bearing amphibolite yielded extensive metamorphic recrystallization age of ca. 530–520 Ma. The reported Hf crustal model ages of the KC are in the range of 1200–2800 Ma (Santosh et al. 2014; He et al. 2015).

**Samples and analytical techniques**

**Sample descriptions and field relations**

Six rock samples (sample numbers: UHT 3C, UHT 4D, UHT 6F, UHT 9I, UHT 10J, and UHT 11K) were collected from three localities in the central HC (Figure 2(a,b)). The details of the samples are given below and their field relations are shown in Figure 2.

**UHT 3C and UHT 4D**

The sampling locality is a quarry in Gampola, which is located slightly close to the ductile shear boundary between the HC and the KC (Kröner, 1991; Voll and Kleinschrodt 1991; Cooray 1994, Figure 1(a,b)). The surrounding area is prominently comprised of garnet-sillimanite-biotite-quartz ± graphite gneisses (Khondalites), quartzites, charnockitic gneisses, biotite-hornblende gneisses, hornblende-biotite gneisses, marbles, and granitic gneisses (Geological Survey and Mines Bureau 1996a, map sheet No. 14). The quarry defines well-foliated rocks with different mineral assemblages as compositional domains. In the rock gradual variation of mineral parageneses without sharp contacts was observed.

Discontinuous layers (approx. 30–40 cm thick) of sapphireine- and kyanite-bearing garnet-sillimanite-orthopyroxene gneiss domains (UHT 3C) occur within the host sillimanite-bearing garnet-orthopyroxene gneiss (UHT 4D) (Figure 2(a,b)). The UHT 3C sample contains mainly medium to coarse subhedral to euhedral garnet (0.25–1.5 cm in diameter; see Supplementary Figure 1(a)) and orthopyroxene...
The rock contains plenty of ribbon quartz (up to 5 cm in length; see Supplementary Figure 1(a)) evidencing strong non-coaxial deformation. Pyroxene is found associated closely with garnet or as isolated grains in the matrix. Feldspar-rich irregular nabs within ribbon quartz in the matrix might indicate that they were crystallized from a silicate melt. Tiny, disseminated biotite flakes are found in the matrix.

The host rock UHT 4D mainly contains porphyroblastic garnet (0.25–2 cm in diameter; see Supplementary Figure 1(b)), porphyroblastic orthopyroxene (up to 1 cm), ribbon quartz (up to 4 cm in length) and feldspars. Occasionally, ribbon quartz, feldspars, and tiny biotite-rich layers (about 1.5 cm thick) are also present within this host rock.

UHT 6F
Garnet and graphite-bearing quartz-feldspathic gneiss was collected from a quarry close to Nawalapitiya (Supplementary Figure 1(c)), adjacent to the tectonic contact between the HC and KC. The surrounding lithologies are strongly deformed garnet-sillimanite-biotite ± graphite gneisses (Khondalites), charnockitic gneisses, quartzites, hornblende and biotite gneisses, marbles and granitic gneisses (Geological Survey and Mines Bureau, 1996a, map sheet No. 14). A detailed description of the quarry is reported in Dharmapiya et al. (2015a).

The UHT 6F occurs as fresh and light-coloured massive bands of about up to ~10 m thickness within the host charnockitic gneisses (Figure 2(c)). In the charnockite
host rock, the UHT 6F occurs parallel to the major foliation of the rock. Quartz-undersaturated domains (corundum- and spinel-bearing garnet-sillimanite–graphite gneisses) frequently occur as patches or boudins with 15 cm to >1 m in diameter (Figure 2(d) and Supplementary Figure 1(d)) within the UHT 6F.

The UHT 6F contains stretched quartz (up to 5 cm in length), recrystallized feldspars, subhedral to euhedral garnet-porphyroblasts (0.25–4 cm in diameter), and fine- to medium-grained graphite flakes. Tiny, acicular biotite flakes occur as thin nabs parallel to the main foliation or around porphyroblastic garnet. Locally occurring biotite, feldspar, and irregular quartz-rich nabs and patches (Supplementary Figure 1(d)) probably have crystallized from a melt.

**UHT 9I, UHT 10J, and UHT 11K**

These three rock samples were collected from a road exposure close to Nildandahinna (Figure 2(e)). The surrounding area is mainly composed of garnet-sillimanite-graphite gneisses (Khondalites), quartzites, charnockitic gneisses, quartzo-feldspathic gneisses, marbles and granitic gneisses (Geological Survey Department of Sri Lanka 1996a, 1996b map sheet Nos. 14 and 17).

More than 90% of the outcrop is composed of foliated charnockitic gneiss. The garnet-orthopyroxene gneiss (UHT 9I) occurs as discontinuous layers (~15–25 cm thickness) in charnockite (Figure 2(e), Supplementary Figure 1(e) and (f)). Sometimes UHT 9I is intimately with spinel- and corundum-bearing garnet-sillimanite-graphite gneiss (Khondalite), mainly found as patches/boudins (up to 0.5 m in diameter, Supplementary Figure 1(e) and (f)) within the charnockite. Details of petrology and geochronology of this corundum-bearing garnet-orthopyroxene gneiss is given in Dharmapiya et al. (2015a). The UHT 9I is comprised of subhedral to anhedral fine to medium grained garnet (<0.2 to 0.75 cm in diameter), fine- to medium-grained orthopyroxene (<0.2 to 0.50 cm), fine- to medium-grained quartz (<0.2 to 0.5 cm), and recrystallized feldspars.

Discontinuous layers of garnet-bearing quartzofeldspathic rock (UHT 10J) (~30 cm thick) cross-cut nearly parallel to the major foliation of the host charnockite (Figure 2(e), Supplementary Figure 1(e) and (g)). This rock contains anhedral porphyroblastic garnet surrounded by feldspars and quartz indicating a peritectic origin (Supplementary Figure 1(h)). Matrix contains relatively large and medium to coarse feldspars (0.25–1 cm) and irregular-shaped quartz.

Garnet-orthopyroxene bearing coarse-grained samples of UHT 11K were collected from melt patches (~20 cm thick) in the contact between host charnockite and granit-orthopyroxene granulate (UHT 9I, Figure 2(n) and (o)). This melt patch clearly cross-cuts the foliation of UHT 9I (Figure 2(f), Supplementary Figure 1(i)). The similar types of melt patches were present within the host charnockite also (Supplementary Figure 1(j)). The UHT 11K contains coarse, anhedral garnet (up to 2 cm), coarse anhedral orthopyroxene (up to 2 cm), and coarse grained and irregular shaped quartz. Biotite at the margin of garnet and orthopyroxene could be the late products due to hydration of the host minerals.

**Analytical technique**

**Mineral chemistry**

Mineral compositions were analysed using a JEOL JXA8530 Field Emission Electron Probe Microanalyzer (FE-EPMA) at the Indian Institute of Science, Bangalore, India. All analyses were carried out using an accelerating voltage of 15 kV and 20 nA beam current with 1–3 μm spot size.

**Zircon U–Pb dating**

After crushing of ~1.5 kg of each rock sample using a jaw crusher, zircon grains were separated by gravimetric and magnetic separation methods. Hand picking under the binocular microscope was carried out at the Centre for Earth Sciences, Indian Institute of Science, India.

Subsequently, the selected grains were mounted in epoxy resin discs and polished to expose mid-sections, before the gains were subjected to gold sputter coating followed by taking images under both transmitted and reflected light. To identify internal structures and choose potential target sites for U–Pb analyses, the zircon grains were imaged under cathodoluminescence (CL) at the State Key Laboratory for Continental Dynamics, Department of Geology, Northwest University, Xi’an, China.

U–Pb isotopes in zircons were determined using a Geolas-193 UV laser ablation system coupled with an Agilent 7500a ICP-MS at the State Key Laboratory of Continental Dynamics, Northwest University in Xi’an. Helium was utilized as the carrier gas with a laser beam of 32 μm in diameter and a frequency of 6 Hz. Data acquisition time was 40 s. Zircon 91500 was used as the standard for U–Pb isotopic ratio determination and age calibration. The standard NIST610 was employed as an external standard and Si as the internal standard during trace element analysis. Common Pb was corrected using measured ²⁰⁶Pb. The degree of discordance of data was <20% during the analysis while in the majority of data with degree of discordance <10%. The detailed analytical procedure is described in Liu et al. (2007). Concordia diagrams and weighted mean calculations were determined using the software Isoplot/Ex Ver3 (Ludwig 2003).
Results

Petrography

A brief description of the textures of the studied rocks is given in the below section. Mineral abbreviations are after Kretz (1983).

**UHT 3C**

In the UHT 3C, sillimanite, quartz, and plagioclase are found as major mineral inclusions in garnet. Biotite and Fe-Ti oxides (exsolved titano-haematite + ilmenite ± rutile) are present as minor minerals while sapphirine, kyanite, rutile, zircon, and monazite occur as accessory phases. Sapphirine and kyanite coexist at the core (Figure 3(a) and, Supplementary Figure 2(a)).

The Supplementary Figure 3 shows the Raman spectrum of the above kyanite coexisting with sapphirine. Presence of biotite, fibrolitic sillimanite, and isolated quartz inclusions from core to rim in most of other garnet indicates prograde reaction, Bt + Sil + Qtz = Grt ± Kfs + Melt (1). Localized orthopyroxene-sillimanite (Figure 3(c)) or biotite-plagioclase coronas indicate the garnet breakdown reactions, Grt + Qtz = Opx + Sil (2) and Grt + Fluid = Bt + Pl (3), respectively.

Ribbon quartz (up to 6 cm), recrystallized plagioclase and K-feldspar are present as major constituents of the matrix (Supplementary Figure 2(b)). Biotite and Fe-Ti oxides (exsolved titano-hematite + ilmenite ± rutile) are found as minor minerals while zircon and monazite are present as accessory.

![Figure 3](image-url)

**Figure 3.** Photomicrographs of the studied Rock samples. (a) Garnet with sapphire, kyanite, quartz, and biotite as major inclusion phases and the coexistence of sapphire and kyanite at the core area of UHT 3C. (b) Sillimanite and quartz inclusions in garnet of UHT 4D. (c) Inclusion phases of garnet in UHT 6F. (d) Inclusion phases in garnet of UHT 9I. (e) The peritectic garnet of UHT 10J. (f) The peritectic garnet of the UHT 11K.
**UHT 4D**
In this rock, quartz, K-feldspar, and plagioclase are present as major inclusion phases in garnet. Sillimanite, spinel, biotite, and Fe-Ti oxides (exsolved titanohematite + ilmenite ± rutile) occur as minor inclusion phases while zircon, monazite, and apatite are the accessory phases (Figure 3(d,e)). The inclusion phases in garnet suggest the prograde reactions (1) and \( \text{Bt} + \text{Sil} = \text{Grt} + \text{Spl} + \text{Melt} \) (4).

Orthopyroxene, biotite, plagioclase, K-feldspar, and quartz occur as major mineral constituents (Figure 3(b) and Supplementary Figure 2(c), (d) and (e)) in the matrix. An Fe-Ti oxide (exsolved titanohematite + ilmenite ± rutile) is the minor constituent. Orthopyroxene often occur close to and/or coexisting with garnet (Supplementary Figure 2(d)). Occasionally garnet is surrounded by biotite-plagioclase symplectite while outer margins of some of the orthopyroxenes are also overprinted by biotite.

**UHT 6F**
In this rock, quartz, plagioclase, and alkali-feldspar are present as major inclusion phases in garnet. Biotite, rutile, and ilmenite are present as minor mineral phases while zircon, apatite, and monazite are found as accessory minerals (Figure 3(c)). These inclusion phases in garnet suggest the prograde reaction: \( \text{Bt} + \text{Qtz} ± \text{Pl} = \text{Grt} + \text{Kfs} + \text{Melt} \) (5).

Less commonly, biotite at the rim of garnet probably indicates the hydration during retrogression via reaction (3). Stretched quartz, strongly recrystallized plagioclase, and alkali-feldspars (Supplementary Figure 3 (f)) occur as major mineral phases in the matrix. Pyrite, ilmenite, and graphite are found as minor mineral phases while zircon and apatite make the set of accessory phases.

**UHT 9I**
In the UHT 9I, biotite, quartz, and K-feldspar are the major inclusion phases while plagioclase (Figure 3(d)), ilmenite and rutile are the minor mineral phases in garnet. Zircon and apatite are present as accessory phases. The above inclusion phases in garnet suggest that the rock has evolved via the prograde melting reaction (5). Locally garnet has broken down forming biotite + plagioclase assemblage probably via the reaction (3).

In the matrix, orthopyroxene, quartz, K-feldspar, and plagioclase are present as major mineral phases (Supplementary Figure 3 (g)). Some of the quartz shows slightly elongated nature. Biotite and ilmenite are present as minor phases while zircon and apatite occur as accessory minerals. Coexistence of garnet–orthopyroxene is commonly observed (Figure 3(d) and Supplementary Figure 3(g)).

**UHT 10J**
Garnets of this rock are mostly inclusion free. Occasionally, quartz and K-feldspar inclusions are present (Figure 3(e)). Due to intergrowths of garnet with matrix quartz and K-feldspar, it can be precluded that these garnets could have formed from a melt as a peritectic phase.

As major matrix minerals, K-feldspar and quartz are present while plagioclase occurs as a minor mineral (Supplementary Figure 3(h)). Some of the quartz grains show elongated form. Medium grained subhedral monazite (up to 0.4 cm), ilmenite and zircon are present as accessory phases.

**UHT 11K**
Garnets of this rock are also mostly inclusion-free and surrounded by quartz and feldspars, probably indicating origin from an existed melt phase, which may have resulted via the prograde melting reaction (reaction 3 and 5) of surrounding rocks (Figure 3(f) and Supplementary Figure 3(i)). Some of the garnets are rimmed with biotite + plagioclase crown (Supplementary Figure 3(i)), indicative of the retrograde reaction (3).

The matrix of this rock is composed of anhedral coarse to medium orthopyroxene, coarse grained plagioclase, and K-feldspar as major mineral phases (Figure 3(f) and Supplementary Figure 3(j)) while biotite is found as a minor mineral phase. Ilmenite, monazite, and zircon are present as accessory phases. Frequently, coarse (up to 3 cm) anti-perthite grains are also present (Supplementary Figure 3(j)).

**Mineral chemistry**
The mineral chemical data are given in Supplementary Tables 1–3 and brief description of mineral compositions is given below.

**Garnet**
Garnets in all the samples are mainly almandine–pyrope solid solutions. Core and rim compositions of the analysed garnet indicate slight compositional zoning of Fe, Mg, and Ca (Supplementary Table 1). The garnet cores are slightly depleted in almandine component whereas pyrope content shows opposite behaviour (Supplementary Table 1). The sample UHT 3C contains pyrope-rich garnet and have uniform pyrope content from core to rim (Grt\(_{\text{py}}\) ~0.54). The garnet in samples UHT 9I, UHT 10J, and UHT 11K are characterized by high grossular content (Grt\(_{\text{Grs}}\) ~ 0.14, 0.12, and 0.15 at core and 0.11, 0.11, and 0.14 at the rim). All the studied garnets show low spessartine contents (Grt\(_{\text{Spe}}\) < 0.03).
**Pyroxene**

The compositions of orthopyroxenes in UHT 3C, UHT 4D, UHT 9I, and UHT 11K are given in the Supplementary Table 2. The $X_{Mg}$ content of orthopyroxene in UHT 3C and UHT 4D are relatively high ($X_{Mg} \sim 0.70–0.72$). However, the $X_{Mg}$ in UHT 9I and UHT 10J are $\sim 0.51$ and $\sim 0.49$, respectively. The orthopyroxene in sample UHT 3C and UHT 4D shows increased amounts of Al $\sim 8.50$ wt% and $7.90$ wt%, respectively. In both samples, orthopyroxene shows slight zonation from core to rim (Al content of cores are $\sim 0.5$ wt% richer, compared to rim).

**Biotite**

The compositional variations of biotite are shown in Supplementary Table 2. Except the biotite inclusions in garnet of UHT 3C, all the other biotites show high TiO$_2$ content (from $4.53$ to $6.80$ wt%). The $X_{Mg}$ content is also relatively higher in biotite inclusions in garnet ($X_{Mg} 0.56–0.81$) compared to biotite in the matrix of all the samples ($X_{Mg} 0.56–0.66$).

**Feldspars**

The compositions of plagioclase and K-feldspar are given in Supplementary Table 3. Plagioclase in all the samples is dominated by albite component (Ab$_{60–80}$). Except for UHT 9I, K-feldspar in all the other samples are rich in orthoclase component (Or$_{96–80}$). The orthoclase content of the UHT 9I is $\sim 0.66$.

**Spinel**

The composition of spinel in the UHT 4D is given in Supplementary Table 3. The $X_{Mg}$ content of spinel is $\sim 0.55$. Cr content of spinel is around $0.10$ wt%. Calculated Fe$^{3+}$ of spinel (based on charge balance) is very low (0.005 pfu).

**Sapphirine**

The Al contents of sapphirine in rock UHT 4D is around 64%. The composition of sapphirine in the UHT 3C is between end member composition of 2:2:1 and 7:9:3. The $X_{Mg}$ content of sapphirine is $\sim 0.74$. Calculated Fe$^{3+}$ of sapphirine (based on charge balance) is very low (0.052 pfu).

**P–T conditions**

The UHT 3C and UHT 4D contain Al-rich orthopyroxene (8.5 wt% and 7.9 wt% of Al$_2$O$_3$, respectively), which could be considered as a typical UHT orthopyroxene (e.g. Harley and Motoyoshi 2000). The Al in orthopyroxene thermometer (Harley and Motoyoshi 2000) for core and rim compositions yielded $\sim 1000–960^\circ$C and $960–920^\circ$C at 10 kbar for orthopyroxene in UHT 3C and UHT 4D, respectively. For UHT 3C, garnet–orthopyroxene geothermometers (mantle compositions of garnet and orthopyroxene) after Lee and Ganguly (1988), Carswell and Harley (1990) and Ganguly et al. (1996) yielded temperatures of 1075–1005°C, 1018–985°C, and 1073–1008°C respectively at 10 kbar while 1067–997°C, 1011–978°C, and 1065–997°C respectively at 9 kbar. However, for the same geothermometers, UHT 4D yielded relatively lower temperatures of 885–858°C, 809–800°C, and 883–849°C at 10 kbar and 879–852°C, 800–794°C, and 877–817°C at 9 kbar, respectively. Dharmapriya et al. (2015b) also reported Al content up to 9 wt% from a porphyroblastic orthopyroxene from sapphirine, kyanite, and spinel bearing garnet–sillimanite–orthopyroxene domain collected from the same quarry. The Al-in-orthopyroxene thermometry yielded peak metamorphic conditions $\sim 1040^\circ$C at 10 kbar. $P$ values calculated using garnet core composition and plagioclase composition using Koziol and Newton (1988) calibration of the GASP geobarometer yielded a range of 7.8–9.8 kbar at $T = 900–1050^\circ$C in the same domain.

Although the rest of the studied rocks lack a typical mineral or mineral assemblage indicating UHT conditions, Dharmapriya et al. (2015b) reported UHT metamorphism from the same localities. Further, Dharmapriya et al. (2015a) calculated the UHT peak metamorphic conditions in the sample of UHT 6F using conventional thermobarometric methods and pseudosection calculations. The authors showed that the rocks in the sampling locality have attained 10–10.5 kbar at 850°C during its prograde evolution. Then the rock has subjected to prograde decompression until the peak metamorphism and the peak metamorphism has been taken place $T$ at 950–975°C and $P$ around 9–9.5 kbar. Subsequently, the rock has undergone near-isobaric cooling stage. During this cooling stage garnet + corundum assemblage was formed via consumption of spinel and sillimanite assemblage, at $T$ $\sim 930^\circ$C at $P$ $\sim 9–9.5$ kbar.

The road exposure from which UHT 9I, UHT 10J, and UHT 11K were collected; Dharmapriya et al. (2015b) calculated $T$ of 900°C using Ti-garnet geothermometer, from garnet of corundum bearing granulite. The peak metamorphic condition of the UHT 9I was calculated using garnet–orthopyroxene geothermometers of Lee and Ganguly (1988), Carswell and Harley (1990) and Ganguly et al. (1996). The thermometers yielded 908–882°C, 810–794°C, and 906–874°C at 9 kbar and 914–889°C, 816–800°C, and 912–880°C at 10 kbar, respectively. The same geothermometers were applied for UHT 11K, which yielded 875–862°C, 787–748°C, and 874–856°C at 9 kbar and 881–868°C, 793–755°C, and 880–862°C at 10 kbar.
P-T calculation data of our recent studies coupled with above P-T calculations (from this study) indicate that the studied rocks could have reached up to 900–950°C at 9–10 kbar during peak metamorphism. Subsequently the rocks have undergone a period of near-isobaric cooling stage, which probably crystallized the in situ melt patches.

**Zircon U–Pb geochronology**

The cathodoluminescence (CL) images of zircons of UHT 3C, UHT 4C, UHT 6F, UHT 9I, UHT 10J, and UHT 11K are shown in Figures 4, 5, and 6, respectively. Zircon U–Pb Concordia plots of rocks UHT 3C and UHT 3C, UHT 4C, and UHT 6F are shown in Figures 7, while those of UHT 9I, UHT 10J, and UHT 11K are shown in the Figure 8.

**UHT 3C**

The zircons of sample UHT 3C are colourless or light brown and transparent to translucent and show various morphologies. The grains show elongated, spherical, near-spherical, or irregular morphology with lengths of 40–140 μm and length-to-width ratios of ~2:1 to ~1:1 in CL images (Figure 4(a)). Most zircons show clear core–rim structures with a dark core, which frequently displays oscillatory zoning, surrounded by bright rim (Figure 4(a)). Structure-less anhedral grains with monotonic grey colour was also found.

A total of 36 spots of 33 zircons were analysed from this rock. The results are shown in Supplementary Table 4. Thirty-five spots are plotted along the Concordia defining seven age groups (Figure 7(a,b)) distinguished as ‘rim ages’ and ‘core ages’. Out of the rim ages, 14 grains gave four tight groups with weighted mean \(^{206}\text{Pb}/^{238}\text{U}\) ages of 536.8 ± 7.4 Ma (MSWD = 0.35, n = 3), 558.4 ± 8 Ma (MSWD = 0.03, n = 2), 603.5 ± 8.7 Ma (MSWD = 2.2, n = 3) and 632.8 ± 5.5 (MSWD = 0.11, n = 3) with ranges of Th/U values of (0.00–0.11), (0.00–0.30), (0.00–0.02), and (0.06–0.24), respectively. The core ages of 21 grains gave four clusters with weighted mean \(^{206}\text{Pb}/^{238}\text{U}\) age of 728.6 ± 6.4 Ma (MSWS = 0.52, n = 3), 837.7 ± 5.6 Ma (MSWD = 0.93, n = 6), 883 ± 5 Ma (MSWD = 0.93, n = 7) and 957 ± 15 Ma (MSWD = 0.00, n = 2) respectively, with Th/U values in the range of (0.35–0.52), (0.22–1.12), (0.33–1.36), and (0.30–0.33), respectively. A single spot gave a concordant age of 1007 ± 13 Ma and another spot yielded a discordant age of 1785 ± 17 Ma.

**UHT 4D**

Zircons in this sample are colourless to light brown and translucent. The anhedral to subhedral grains show elongated, near-spherical, or irregular forms. The length
of the grains varies from 40 to 170 μm and length to width ratios varies from 2.5:1 to 1:1. Large numbers of most zircons show clear core–rim texture in CL images (Figure 4(b)). Zircon cores show well preserved oscillatory zoning.

A total of 35 spots of 33 zircons were analysed. The results are shown in Supplementary Table 4. Thirty-one spots yielded concordant ages defining seven age clusters from 524 ± 3.5 Ma to 985 ± 10 Ma. Rim ages were defined by 12 grains in three tight groups (Figure 7(c,d)) with weighted mean $^{206}$Pb/$^{238}$U ages of 534 ± 10 Ma (MSWD = 0.08, $n = 2$), 556.6 ± 3.6 Ma (MSWD = 0.23, $n = 7$), and 607 ± 7 Ma (MSWD = 0.53, $n = 2$). The Th/U values of these clusters were (0.02–0.18), (0.01–0.06), and (0.01–0.06), respectively. The core age of 19 grains of rest of the 21 grains gave four clusters with weighted mean

Figure 5. Cathodoluminescence (CL) images of representative zircons in UHT 6F.

Figure 6. Cathodoluminescence (CL) images of representative zircons in (a) UHT 9I, (b) UHT 10J, (c) UHT 11K.
$^{206}\text{Pb}/^{238}\text{U}$ ages of $733.5 \pm 5.2$ Ma (MSWS = 0.13, $n = 3$), $863.4 \pm 6.9$ Ma (MSWD = 0.95, $n = 3$), $931 \pm 8.2$ Ma (MSWD = 1.2, $n = 7$), and $979 \pm 10$ Ma (MSWD = 0.28, $n = 4$), respectively. The Th/U values of above clusters were in the ranges of (0.13–0.37), (0.44–3.61), (0.36–0.78), and (0.32–0.65), respectively. Three spots gave discordant ages from $1152.6 \pm 8$ Ma to $1574.7 \pm 11$ Ma.

**UHT 6E**

The zircons from this sample are colourless and transparent to translucent. The anhedral to subhedral grains show near-spherical or irregular morphology. The length of grains is in the range of 25–140 μm and length-to-width ratios varying from 2:1 to 1:1. Most zircons show clear dark core and bright rim in CL...
images (Figure (5)). Some of the cores contain oscillatory zoning. Occasionally, structure-less discrete grains are present with homogeneous grey colour.

A total of 26 spots of 24 zircons were analysed (Figure 5). The results are shown in Supplementary Table 4. Seventeen spots yielded concordant ages defining two clusters in which seven spots yielded a tight group of weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 532.9 ± 3 Ma, (MSWD = 0.74, $n = 7$; Figure 7(e,f)) with a wide range of Th/U (0.02–0.57). There are 10 spots yielded tight group of weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 554.7 ± 2.7 Ma (MSWD = 0.19, $n = 10$). There was a single grain giving $^{206}\text{Pb}/^{238}\text{U}$ concordant age of 761 ± 10 Ma with Th/U value 1.20. Rest of the spots yielded discordant ages from 807.6 ± 7 to 1872.2 ± 15 Ma (Figure 7(e)).

**UHT 9I**

The zircons from sample UHT 9I are colourless to light brownish and transparent to translucent. The grains show spherical, near-spherical, or irregular morphology. The length of zircon varies from 40 to 160 μm while the length-to-width ratio varies from 2:1 to 1:1. Most zircons show clear core–rim texture in CL images (Figure 6(a)). Frequently, the grains show dark core and bright rims. Occasionally, discrete grains with uniform grey colour are present.

A total of 24 spots of 24 zircons were analysed. The results are shown in Supplementary Table 4. The cores of three analysed grains yielded concordant $^{206}\text{Pb}/^{238}\text{U}$ ages of 2164 ± 18 Ma, 2573 ± 16, and 2862 ± 17 Ma (Figure 8(a)). Their Th/U ratios are 1.55, 0.53, and 0.58, respectively. Another three groups yielded concordant weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 575.8 ± 6.4 Ma (MSWD = 0.80, $n = 2$), 611.6 ± 6.7 Ma (MSWD = 1.12, $n = 3$), and 665 ± 5.9 Ma (MSWD = 0.06, $n = 3$), respectively (Figure 8(a,b)). Their Th/U ratios range 0.03–0.05, 0.03–0.07, and 0.02–0.06, respectively.

**UHT 10J**

Zircons in sample UHT 10J are colourless and transparent to translucent. The subhedral to anhedral grains show near-spherical or irregular morphology. The lengths of the grains vary from 30 to 190 μm and the length-to-width ratios vary in the range of 2:1 to 1:1. Most zircons show clear core–rim texture in CL images (Figure 6(b)). Few grains display uniform grey colour without any specific structure.

Twenty-four spots of 24 zircons were analysed. The results are shown in Supplementary Table 4. All the points define concordant behaviour with ages ranging from 534 ± 4 Ma to 567 ± 4 Ma as two tight groups (Figure 8(c)). Twenty-one spots defined a tight group...
yielding a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 557.9 ± 1.7 Ma (MSWD = 0.89) while the other group yielded a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 536.3 ± 4 Ma (MSWD = 0.40). The Th/U ratios of zircon yielded relatively higher values in ranges of 0.66 – 0.78 and 0.50 – 0.91.

**UHT 11K**

Zircons in sample UHT 11K are colourless and transparent showing mainly an irregular morphology. The length-to-width ratios of zircons range from 30 to 180 µm. Most of the zircon grains were structure-less with homogeneous grey colour in CL images (Figure 6(c)).

A total of 35 spots of 35 zircons were analysed. The results are shown in Supplementary Table 4. The analysed grains define two tightly weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age groups (Figure 8(d)). One group of 11 spots yielded weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 533.9 ± 2.4 Ma (MSWD = 0.40) and the other group (24 spots) defined a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 552.6 ± 1.6 Ma (MSWD = 0.87). The Th/U ratios of the two groups were in the ranges of 0.19–0.62 and 0.33–0.90, respectively.

**Discussion**

**Protolith, metamorphic, and in situ melt crystallization ages**

Zircons in the UHT 3C define four populations of concordant ages (Figure 7(a,b)) from their dark cores (with well-preserved oscillatory zoning, Figure 4(a)), of $^{206}\text{Pb}/^{238}\text{U}$ age (weighted mean) 728.6 ± 6.4, 837.7 ± 5.6, 883 ± 5, and 957 ± 15 Ma, together with a single grain defining a $^{206}\text{Pb}/^{238}\text{U}$ age of 1007 ± 13 Ma. The ages of these detrital grains reflect the timing of magmatic crystallization in their protoliths. The light colour anhedral grains and light colour rims of darker grains with low Th/U ratios (Th/U = 0.00–0.11, except one spot) are consistent with typical metamorphic origin and could be categorized into four weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age groups as, 632.8 ± 5.5 Ma, 603.5 ± 8.7 Ma, 558.4 ± 8 Ma, and 536.8 ± 7.4 Ma. These age clusters clearly represent the time span of the metamorphism. The single discordant inherited age of 1785 ± 16 Ma may indicate incorporation of sediments that derived from probably Palaeoproterozoic sources.

The sample UHT 4D, collected from the same quarry as the sample UHT 3C was collected also yield four populations of concordant weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages (Figure 7(c,d)) from their dark cores with well-preserved oscillatory zoning (Figure 7(a)), of 733.5 ± 5.2 Ma, 863.4 ± 6.9 Ma, 931.2 ± 8.2 Ma, and 979 ± 10 Ma. In addition, three single grains define $^{206}\text{Pb}/^{238}\text{U}$ ages of 778 ± 6 Ma, 807 ± 9 Ma, and 899 ± 6 Ma. All these ages reflect the timing of crystallization from the magmatic protoliths of these detrital grains. The zircon rims with low Th/U ratios (except one grain, all others <0.06) indicate that they are of typical metamorphic origin and yield three coherent tight groups with weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 607 ± 7 Ma, 556.6 ± 3.6 Ma, and 534 ± 10 Ma. The oldest discordant inherited age of 1574.7 ± 11 Ma give a clue for contributions of Palaeoproterozoic protoliths sediments.

Dharmapriya et al. (2015b) measured LA-ICPMS zircon U–Pb ages of a sapphire, kyanite, and spinel bearing garnet–sillimanite–orthopyroxene domain of the same quarry from which the UHT 3C and UHT 4D were collected. The authors reported two populations of concordant ages from dark-zircon cores of 834 ± 12 Ma and 722 ± 14 Ma, reflecting the age of magmatic crystallization of the protoliths, while another coherent group yielded a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 535.2 ± 4.8 Ma (MSWD = 0.22, n = 18) representing the metamorphism.

Zircons in the UHT 6F contain low Th/U ratio (mostly <0.10) representing two coherent groups (Figure 5(a,b)) with weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 554.7 ± 2.7 Ma and 532.9 ± 3 Ma, reflect the metamorphic origin. A single grain with $^{206}\text{Pb}/^{238}\text{U}$ core age of 761.8 ± 10 M with high Th/U ratio (1.20), probably represents a magmatic zircon derived from Neoproterozoic protolith. The detrital zircons in this sample showed discordant ages from 808 to 1873 Ma.

Dharmapriya et al. (2015b) measured LA-ICPMS zircon U–Pb ages of the quartz-undersaturated domains (corundum- and spinel-bearing garnet-sillimanite–graphite gneisses) of the same quarry from which the UHT 6F was collected, and reported weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 530.2 ± 3.7 Ma as the metamorphic age (MSWD = 0.78, with a wide range of Th/U from 0.02 to 0.57). The detrital zircons of this study yielded inherited ages between 1676 and 1787 Ma with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 1722 ± 31 Ma (MSWD = 0.6, N = 7, Th/U = 0.07–0.82) along with a concordant age of 759 ± 13 Ma (Th/U = 0.04) defined by a single grain. Hence our data indicate that the studied metasediments in the quarry represent majority of Palaeoproterozoic and Neoproterozoic protolith materials.

Zircons of the UHT 9I form three coherent groups with concordant ages of 575.8 ± 6.4 Ma, 611.6 ± 6.7 Ma, and 665.5 ± 5.9 Ma with low Th/U ratios (less than 0.06) reflecting a metamorphic origin. Magmatic zircons with clear oscillatory-zoned core yield discordant ages from 736 to 2303 Ma and concordant ages of 2164.1 ± 18 Ma, 2454.9 ± 15 Ma, 2573.4 ± 16 Ma, and 2862.2 ± 17 Ma inferring the detrital old ages represent Palaeoproterozoic to Neoarchaean protolith materials as their sources.
Dharmapriya et al. (2015b) measured the U–Pb ages of corundum bearing granulite collected from the same outcrop from which UHT 9I, UHT 10J, and UHT 11K were collected, and presented a coherent group of $^{206}\text{Pb}/^{238}\text{U}$ ages from metamorphic zircons with a weighted mean age of $578.0 \pm 3.7$ Ma (MSWD = 0.35, $N = 19$) with low Th/U ratios (up to 0.16) reflecting typical metamorphic origin. The $^{206}\text{Pb}/^{238}\text{U}$ age of the core in two zircons of this study yielded an age of $2365 \pm 31$ Ma (Th/U = 1.33) reflecting Palaeoproterozoic contributions.

The sample UHT 10J yielded two coherent groups of $^{206}\text{Pb}/^{238}\text{U}$ ages of $558 \pm 1.6$ Ma and $536.3 \pm 4.4$ Ma. Th/U ratios of both groups are 0.50–0.81 and 0.66–0.78, respectively, thus they deviate from the typical metamorphic origin. The rock layer occurs as metre scale discontinuous layers (Figure 2(e), Supplementary Figure 2j) or irregular patches more or less crosscutting the major foliation showing it as crystallized from a melt phase after the peak metamorphism.

The UHT 11K could also have derived from crystallization of in situ melt, which probably derived from partial melting of host charnockite during the late pro-grade to peak metamorphism. This rock also crosscuts the major foliation of the UHT 9I and host charnockite (Figure 2(f)). The rock also yielded two coherent groups of $^{206}\text{Pb}/^{238}\text{U}$ ages, $552.6 \pm 1.6$ and $533.9 \pm 2.4$ Ma. Th/U ratios of the two groups are 0.34–0.90 and 0.36–0.62 (except one grain) clearly indicating magmatic origin.

Our data show crystallization of in situ melt at Nildandahinna (UHT 10J and 11K) has taken place from ca. 558–535 Ma. Tsunogae et al. (2014) have argued that the dissolution of old zircons during partial melting and regrowth of new zircon with coexisting U- and Th-bearing minerals can allow the growth of high Th/U zircon. Dharmapriya et al. (2015b) presented $^{206}\text{Pb}/^{238}\text{U}$ ages of $542 \pm 2$ Ma (peak metamorphism) from the UHT khondalite at Gampola and Nawalapitiya (close to the present sampling localities at the KC–HC boundary) and a cooling age of $514 \pm 3$ Ma. Possibly the latter age may represent the timing of melt crystallization in the central HC. Hence it appears that a time span of ca. 25–30 Ma has lapsed for the melt crystallization between the eastern end of the HC (e.g. Nildandahinna) and the central HC (e.g. Kotmale).

**Implications for the depositional history of the HC metasediments**

Deposition of sediments of the HC took place from 3200 Ma to 1900 Ma (e.g. Baur et al. 1991; Hö1zl et al. 1991, 1994; Kröner et al. 1994). Intrusions of most of the granitoid plutons into the HC sediments have taken place at 1800–1900 Ma and very rarely magmatic activities are recorded later than 1800 Ma except for a single granitic intrusion at ca. 670 (Baur et al. 1991) and mafic intrusion at ca. 920 Ma (Dharmapriya et al. 2015b). This long silent period without a significant tectono-magmatic activity in the HC was followed by pervasive thermal events of regional granulite facies metamorphism during the Ediacaran period (between 610 and 550 Ma) in response to collision of HC and WC associated with the assembly of Gondwana Supercontinent (e.g. Hö1zl et al. 1991, 1994; Kröner et al. 1994). Accordingly, there has been ca. 1500 Ma time span between the last deposition of HC sediments and its high-grade metamorphism without sufficient evidence for magmatic activity, however, remains unexplained (e.g. Perera and Kagami 2011).

Our Palaeoproterozoic to Neoarchaean detrital zircon ages from the sample UHT 9I in this study clearly reflect the ages of magmatic protolith materials of the sample, which is well consistent with the ages derived by previous studies. However, the new age data of the present study (UHT 3C and UHT 4D) and Dharmapriya et al. (2015b), we reveal that the minimum depositional age of the protolith sediments are as young as 720 Ma. Further, incorporation of zircon grains with discordant $^{206}\text{Pb}/^{238}\text{U}$ inherited age of 1575 Ma to 1873 Ma in these two localities precisely indicate contributions from Palaeoproterozoic sources. Previous reports of relict detrital cores with SHRIMP ages of ca. 2500–830 Ma and clusters of ca. 1700 and 1040–830 Ma representing episodes of zircon growth in the Palaeo–Neo Proterozoic protolith sediments (e.g. Sajeev et al. 2010) further support our interpretations.

In Sri Lanka, the oldest inherited age known so far (3200 Ma) was reported by Kröner et al. (1987) from zircons of a metapelite sample close to Polonnaruwa. Subsequently, Amarasinghe and Collins (2011) presented 90% concordant $^{207}\text{Pb}/^{206}\text{Pb}$ ages of the detrital zircons from quartzite of the HC with major detrital peaks at 2200, 2500, and 2700 Ma. Therefore, incorporation of inherited zircons with $^{206}\text{Pb}/^{238}\text{U}$ concordant ages up to 2862 Ma in UHT 9I in this study indicates the extension of the protolith age of the studied rocks into the Neoarchaean. Hence, the new ages of detrital zircon populations of this study imply that shallow marine sediments have derived from Neoarchaean to Palaeoproterozoic multiple age provenances.

In the HC, the U–Pb Concordia upper intercept ages of 1950–1850 Ma of some orthogneisses have been interpreted as the intrusion age of their parent magma (Baur et al. 1991; Hö1zl et al. 1991, 1994; Kröner et al. 1994; Santosh et al. 2014). However, we prefer to interpret these ca. 2000–1800 Ma magmatic ages serve best to provide the evidence for existence of a
Palaeoproterozoic to Neoproterozoic basement as a platform for deposition of the HC marine sediments. The Ediacaran–Cambrian metamorphic event may have caused the early reworked older basement rocks to be repetitively reworked subsequently.

Our interpretations are further supported by published U–Pb and Lu–Hf isotope data from the HC. Milisenda et al. (1994) interpreted that metagneous and metasedimentary rocks in the HC have pre-600 Ma crustal history and represent extensive reworking of old continental crust. Santosh et al. (2014) reported Lu–Hf data of zircon of mafic to intermediate granulites and charnockites revealing that the HC rocks preserve distinct imprints of reworking of older crust. The zircon εHf(t) values in their mafic granulite displayed a tight cluster from −2.2 to 0.1 with Hf crustal model ages ($T_{DM}^S$) in the range of 1501–1651 Ma suggesting a mixed source from both juvenile Neoproterozoic and reworked Mesoproterozoic components. Zircons in their metagabbro showed negative εHf(t) values (mean −6.3) and depleted mantle model ages ($T_{DM}$) of 1847–1978 Ma suggesting reworked Palaeoproterozoic crust as the source. Zircons in massive charnockites and metadiorites displayed highly negative εHf(t) values and older Hf crustal model ages (up to 2790 Ma) suggesting reworked Neorchaean–Palaeoproterozoic crust. In addition, He et al. (2015) reported that zircons in a charnockite with crystallization age of 565 Ma in the HC shows negative εHf(t) values in the range of −6.7 to −12.6 with Hf crustal model ages of 2039–2306 Ma suggesting magma derivation through melting of a Palaeoproterozoic source. Also, a metadiorite sample (crystallization age ~ 576 Ma) in the same sampling locality showed εHf(t) range from −11.1 to 1.6 suggesting a mixed source of both older crustal and juvenile material.

Hence, the Palaeoproterozoic to Neoproterozoic igneous basement may have provided the platform for deposition of the HC sediments derived from multiple sources. The Neoproterozoic collisional event around Sri Lanka has resulted in intense reworking of the older crust. The magma derived from melting of the already reworked crust may have given rise to crystallization of granitic, charnockitic and metadioritic rocks with diverse age populations. Subsequently, the remnants of the existed basement rocks (residue after melting of the already reworked crust) could have metamorphosed under granulite facies during the Ediacaran–Cambrian thermal event. The geochemical studies of the HC metasediments indicating that the deposition of sediments has taken place in a stable shelf region of shallow marine environment (e.g. Dissanayake and Munasinghe 1984; Kröner et al. 1994; Prame and Pohl 1994; Santosh et al. 2014) further support above arguments.

Late Cryogenian to early Cambrian multiple thermal events of the Highland Complex and the neighbouring Gondwana terrains

As summarized in previous sections, several metamorphic ages have been proposed for the Sri Lankan basement. From Rb–Sr data the age of metamorphism was considered as ca. 2000–2500 Ma (e.g. Crawford and Oliver 1969; De Maesschalck et al. 1990; Kagami et al. 1990). Using U–Pb zircon ages Kröner et al. (1987) earlier, argued that the regional metamorphic event of the HC to be at ca. 1100 Ma. However, there are number of studies showing U–Pb zircon ages confining the age of metamorphism at ca. 610–550 Ma simultaneous with the assembly of Gondwana (Baur et al. 1991; Hö1zl et al. 1991, 1994; Kröner and Williams 1993; Kröner et al. 1994, 2003, 2013; Sajeev et al. 2010). Further, recent studies from the HC further expanded the duration of the time span of this late Neoproterozoic–Cambrian event by providing evidence for multiple thermal activities from 728 to 511 Ma (e.g. Malaviarachchi and Takasu 2011b; Santosh et al. 2014; He et al. 2015; Takamura et al. 2015). This Edicaran–Cambrian high-grade metamorphism has been geochronologically well documented in all the neighbouring Gondwana terrains such as southern and central Madagascar (Paquette et al. 1994; Collin et al. 2012), Southern Granulite Terrain of India (e.g. Collins et al. 2007b; Clark et al. 2009a, 2009b; Brandt et al. 2011), and East Antarctica (e.g. Black et al. 1992; Shiraishi et al. 1994; Tsunogae et al. 2015) strongly suggesting terrain amalgamation during the Neoproterozoic.

The weighted mean $^{206}\text{Pb}/^{238}\text{U}$ metamorphic ages obtained by our samples are 665.5 ± 5.9 Ma, 611.6 ± 6.7 Ma, and 575.8 ± 6.4 Ma, from UHT 9I, 632.8 ± 5.5 Ma, 603.5 ± 8.7 Ma, 558.4 ± 8 Ma, and 536.8 ± 7.4 Ma from the UHT 3C and 607 ± 7 Ma, 556.6 ± 3.6 Ma, and 534 ± 10 Ma from the UHT 4D. All these ages are representative of multiple thermal events suffered by the UHT granulites of the HC from late Cryogenian to early Cambrian time. The geochronological evidence for the Ediacaran–Cambrian multiple metamorphism from 612 to 534 Ma is well consistent with the reported previous U–Pb metamorphic zircon ages from the Sri Lankan UHT granulites (Sajeev et al. 2007, 2010; Dharmapriya et al. 2015b) and HT ordinary granulites (Baur et al. 1991; Hö1zl et al. 1991, 1994; Kröner and Williams 1993; Kröner et al. 1994; Santosh et al. 2014; He et al. 2015; Takamura et al. 2015) in the HC. Dharmapriya et al. (2015b) interpreted that the time span of UHT metamorphism in the HC is from 580 to 530 Ma. The sharp peak around 580–530 Ma in the compiled age data probability curve (Supplementary
Figure 4) of studied samples also further confirm that the late Neoproterozoic UHT metamorphism of the HC. However, our older metamorphic zircons with weighted mean $^{206}\text{Pb} / ^{238}\text{U}$ age of 665.5 ± 5.9 Ma and 632.8 ± 5.5 Ma propose a new evidence for early prograde metamorphism of the HC.

Kriegsman, (1993, 1996) and Dharmapriya et al. (2014a) argued that the HC underwent low pressure, relatively high temperature amphibolite facies metamorphism prior to crustal thickening. Kriegsman (1993) further argued that this low-pressure high-temperature metamorphism could be due to crustal extension. Baur et al. (1991) reported evidence for 670 Ma intrusions of granitoids in the HC. Malaviarachchi and Takasu (2011b) also presented CHIME dating of monazite yielding a wide age range of ~728–460 Ma as a new metamorphic age group for the metapelites of Sri Lanka. Sajeev et al. (2007) showed U–Pb concordant ages from zircon cores up to ca. 661 Ma in a mafic granulite collected from the central HC. Recently, Santosh et al. (2014) presented U–Pb concordant ages of ca. 665 Ma from magmatic zircons in a mafic sill within metasediments of the HC. Hence our early metamorphic ages (665.5 ± 5.9 to 534 ± 10 Ma) could represent the late Cryogenian granitic and mafic magmatism associated with early prograde metamorphism in the HC.

Consideration of East Gondwanan terrains, South Kenya, and Tanzania, indicates a major episode of high-grade metamorphism at 640 Ma related with the East African orogeny (Meert 2003). It was suggested earlier that lack of evidence for ca. 640 Ma orogenetic episode in Sri Lanka, East Antarctica, and India could be due to late collision of these terrains and hence unaffected by the older deformation (Meert 2003). However, recent studies of Gondwana terrains, which were adjacent to Sri Lanka, reveal rare evidence for ca. 670–620 Ma metamorphism (e.g. Brandt et al. 2011; Sengupta et al. 2015; Johnson et al. 2015). Brandt et al. (2011) showed U–Th–total Pb old monazite ages from 656 ± 65 to 620 ± 59 Ma in Panhil Hills, Madurai block, Southern India, Sengupta et al. (2015) reported new zircon growth from 633 to 487 Ma from a metamellite in alkaline granite Complex of Sankagiri, Southern granulite terrain, India and Johnson et al. (2015) presented older metamorphic age ca. 640 Ma from metasediments in the Nagercoil Block of Southern India. Additionally, Paquette and Nédélec (1998), showed that Madagascar collided with East Africa at about 650 Ma and underwent extensional collapse at ~630 Ma. Collins (2006) argued that the Antananarivo Block in Madagascar was thermally and structurally reworked between 700 and 532 Ma and during this reworking process, pre-existing rocks were subjected to granulite facies metamorphism. Further, Ashwal et al. (1999) argued that a series of amphibolite facies metamorphic events occurred in the Southern Madagascar between 630 and 530 Ma. De Wit et al. (2001) provided more evidence for this argument based on structural and geochronological studies. Shiraiishi et al. (2008) reported evidence for granulite–facies metamorphism at ca. 600–650 Ma in the Northeastern terrene of Sør Rondane Mountains in the Eastern Drowning Maud Land, East Antarctica. They suggested that theses rocks have undergone amphibolite facies metamorphism at ca. 570 Ma. In contrast, Nakano et al. (2013) showed a 655–635 Ma thermal event from the Sør Rondane Mountains related with granulite-facies metamorphism. Thus, late Cryogenian to early Cambrian multiple thermal events of Sri Lanka and its neighbours clearly indicate existence of robust geochronological evidence to correlate the adjacent terrains.

**Geochronological correlation of Sri Lanka and Southern Granulite Terrain of India**

Southern India is composed of a collage of polymetamorphic terrains with prolonged crustal evolution history from early Archaean to late Neoproterozoic (e.g. Santosh et al. 2006; Santosh and Sajeev 2006). Based on protolith origins and tectonothermal histories, Southern India has been divided into a series of tectonic units (Supplementary Figure 5) from north to south, called: the Salem Block (SB, the Northern Granulite Terrene); the Palghat–Cauvery shear zone (PCSS); the Madurai Block; Achankovil Shear Zone (ACSZ); the Trivandrum Block (the Kerala Khondalite Belt); and the Nagercoil Block (e.g. Gosh et al. 2004; Santosh and Sajeev 2006; Santosh et al. 2009; Collins et al. 2015). The Northern Granulite Terrain is separated by the east–west striking crustal-scale Palghat–Cauvery Shear Zone (e.g. Santosh and Collins 2003a), recognized from the dominantly metasedimentary gneisses and granulites that experienced high-grade metamorphism and pervasive deformation in Neoproterozoic–Palaeozoic times. This southern region is known as the Southern Granulite Terrain (e.g. Santosh and Sajeev 2006).

Approximately 70 km × 400 km east–west extending Palghat–Cauvery Shear zone System (PCSS, Supplementary Figure 5) is characterized by a network of mainly dextral shear zones, located on the southern margin of the Salem Block (Supplementary Figure 5) (Chetty et al. 2003; Tomson et al. 2006; Sato et al. 2011b; Collins et al. 2015). The Madurai Block, which is the largest crustal block in the Southern Granulite Terrain (SGT) and geochronologically, has itself been subdivided into separate terrains as Northern Madurai Block (NMB) and the Southern Madurai...
Block (SMB, e.g. Plavsa et al. 2012, 2014; Collins et al. 2015). The Karur–Kamban-Painavu–Trichur lineament (KKPT) is the lithological boundary that separates the Neoarchean northwestern Madurai Block from the rest of the predominantly metasedimentary Madurai Block (Ghosh et al. 2004). Some workers (e.g. Plavsa et al. 2012, 2014, 2015; Collins et al. 2015) defined an isotopic boundary between Neoarchean to Palaeoproterozoic NMB and Meso- to Neoproterozoic SMB.

The Achankovil Shear Zone (ACSZ) demarcates the southern limit of the Madurai Block, as well as the northern limit of the Trivandin Block and has traditionally been considered as a typical shear zone (Drury et al. 1984; Sacks et al. 1997; Rajesh et al. 1998; Tadokoro et al. 2008). Based on geophysical surveys across the region, some workers have suggested that the ACSZ is a terraine suture (Santosh et al. 2009; Dhanunjaya Naidu et al. 2011). The southernmost tip of India, which is mainly composed of massive charnockites, has been described by some authors as an isolated tectonic unit called the Nagercoil Block (NB) (Santosh 1996; Santosh et al. 2003b).

Previous workers attempted to correlate the SGT of India and the HC of Sri Lanka based on lithologies (e.g. Fernando 1949; Pichamuthu 1967; Yoshida 1988; Braun and Kriegsman 2003), geological structures and metamorphic grades (e.g. Katz 1974, 1978; Braun and Kriegsman 2003), mineralization trends (e.g. Dissanayake and Chandrajith 1999), and geochronological aspects (e.g. Crawford and Oliver 1969; Braun and Kriegsman 2003; Teale et al. 2011). However, lack of sufficient geochronological data was a drawback for appropriate correlation between the two terrains. During the last decade, a number of geochronological studies revealed new magmatic, sediment deposition and crustal evolution ages in the SGT (e.g. Cenki et al. 2004; Ghosh et al. 2004; Santosh et al. 2006; Collins et al. 2007b; Collins et al. 2015; Teale et al. 2011; Kröner et al. 2012; Plavsa et al. 2012, 2014) and the HC (e.g. Sajeev et al. 2007, 2010; Malaviarachchi and Takasu 2011b; Santosh et al. 2014; He et al. 2015; Takamura et al. 2015; Dharmapriya et al. 2015b) presenting more supportive evidence to trace Gondwanian linkage of Sri Lanka with Southern India. The Figure 9 summarizes the data of crust formation ages (Nd model ages and Hf model ages), magmatic zircon ages, detrital zircon/monazite ages, and metamorphic zircon/monazite ages of the HC and the SGT.

### Correlations of crust formation ages of Sri Lanka and India

Nd model ages for the SGT are in the range of 1200–3500 Ma (e.g. Choudhary et al. 1992; Harris et al. 1994; Jayananda et al. 1995; Meißner et al. 2002; Cenki et al. 2004; Tomson et al. 2013). The SB represents Nd model ages in the range of 2500–3500 Ma (e.g. Bhaskar Rao et al. 1996; Bartlett et al. 1998; Tomson et al. 2006; Thomson et al. 2013). The Nd model ages of NMB vary from 2400 to 3200 Ma (Bhaskar Rao et al. 1996; Bartlett et al. 1998; Tomson et al. 2006, 2013) whereas 1200–2800 Ma in the SMB (e.g. Plavsa et al. 2012; Tomson et al. 2013).

Although, the Nd model ages of the ACSZ are initially considered to be younger (between 1200–1600 Ma, e.g. Harris et al. 1994; Brandon and Meen 1995; Bartlett et al. 1998; Cenki et al. 2004), Tomson et al. (2013) reported a range from 1200 to 2800 Ma. The Nd model age of the TB and NB are 2000–2800 Ma (e.g. Cenki et al. 2004; Tomson et al. 2013) and 2200–2800 Ma (Cenki et al. 2004; Tomson et al. 2013).

There are a limited number of Hf crustal model ages the have been reported from both Sri Lanka and SGT of India (Figure 9). The Hf model ages of the SB fall between 2700 and 2930 Ma and $\varepsilon$Hf(t) between +0.5 and +9.2, indicating the detritus was derived from relatively juvenile Archaean terranes. The NMB yield a wide range of Hf crustal model ages from ca. 2500–3900 Ma ($\varepsilon$Hf(t) from −36.9 to +8.7) indicating reworking of Archaean crust with possibly mixing with some juvenile material. The Hf crust formation ages of the SMB range between 1.30 and 3.65 Ga and $\varepsilon$Hf(t) between −31.2 and +7.7 while most of the juvenile signatures ($\varepsilon$Hf(t) ranging from +3.8 to +5.8) are reported from 1000–1100 Ma detrital grains in metasediments. Kröner et al. (2012) reported Hf crustal model ages ranging from 2680 to 3370 Ma ($\varepsilon$Hf(t) between −6.1 and −9.2) from a charnockite which define 207Pb/206Pb upper intercept ages at 1881 ± 14 Ma from samples of TB indicating reworking of Neoarchean to Palaeoarchean crust. The reported Hf crustal model ages of the ACSZ are in the range of 2950–3140 Ma ($\varepsilon$Hf(t) between −6.0 and −17.0) which indicate the reworking of Mesoarchean crust.

The Hf crustal model ages of the HC are in the range of 1500–2800 Ma (Santosh et al. 2014; He et al. 2015). The obtained ages from mafic granulite ca. 1500–1650 Ma and interpreted as mixed source from both juvenile Neoproterozoic and reworked Mesoproterozoic components (Santosh et al. 2014; He et al. 2015). The Hf crustal model ages are in the range of 1850–2800 Ma of metababby, mafic granulites, and charnockites, and have been interpreted as the age of source material for the magma evolved from reworking of the Neoarchean–Palaeoproterozoic crust. Santosh et al. (2014) reported Hf crustal model ages of 2156–3580 Ma (mean 2614 Ma and $\varepsilon$Hf(t) between −33.3 to −10.5) for charnockite sample with upper intercept age of 1812 ± 63 Ma, well correlate with the reported upper intercept ages and Hf model ages (Kröner et al. 2012) of the charnockite in the TB.
Correlations of magmatic ages

Geochronological studies on SB, PCSS, and NMB indicate that the sedimentation of the protoliths of the metasedimentary rocks have taken place on the Archaean/Palaeoproterozoic gneissic basement over a prolonged period of time (e.g. Ghosh et al. 2004; Clark et al. 2009b; Brandt et al. 2011; Sato et al. 2011a; Plavsa et al. 2012; Yellappa et al. 2012; Sengupta et al. 2015). The 820–750 Ma granites, tonalites, and gabbros intruding in to ca. 2500 Ma basement and supra-crustal rocks have been reported in the NMB (e.g Kooijman et al. 2011; Teale et al. 2011; Plavsa et al. 2015) and PCSZ (e.g Sato et al. 2011b; Santosh et al. 2012).

However, there is no evidence so far reported for existence of Archaean/Palaeoproterozoic gneissic basement in SMB (e.g Collins et al. 2015) which contains evidence for Neoproterozoic (1000–790 Ma) magmatism (Plavsa et al. 2012; Sato et al. 2012; George et al. 2015). Evidence for Palaeoproterozoic and Mesoproterozoic (1500–950 Ma) magnetism has been reported in ACSZ (Kröner et al. 2012; Collins et al. 2015; Plavsa et al. 2015). The TB contains evidence for ca. 1800 Ma (Kröner et al. 2012) and ca. 950 Ma (Ghosh et al. 2004) magmatism while the NB reported evidence for ca. 2000 Ma (Ghosh et al. 2004; Collins et al. 2015) magmatism.
The 1800–2000 Ma magmatism in the HC is coeval with the Palaeoproterozoic magmatism in the TB and NC (e.g. Kröner et al. 2003). Dharmapiya et al. (2015b) reported rare evidence for ca. 920 Ma magmatism from the HC. In contrast, there is no evidence for 670–650 Ma magmatism in the TB and the NB reported so far, although it has been reported in the HC (Figure 9). However, the WC of Sri Lanka shows Neoproterozoic magmatism from 1000–730 Ma (Hörlzl et al. 1991, 1994; Santosh et al. 2014; He et al. 2015) fitting with the magmatism in the SMB (e.g. Teale et al. M. 2011; Collins et al. 2015).

**Correlation of detrital zircon/monazite ages**

Braun and Kriegsman (2003) discussed that although Sm–Nd fractionation during the partial melting may have occurred, these distinct Nd isotope systematics suggest derivation of the sediment from Palaeo-, Meso- and Neoproterozoic sources in SGT. However, later studies (e.g. Santosh et al. 2003b, 2006; Collins et al. 2007b, 2015; Kooijman et al. 2011; Plavsa et al. 2014, 2015; Sengupta et al. 2015) revealed the incorporation of Neoarchean to Neoproterozoic sources in the metasediments of the SGT.

The detrital zircon spectra of NMB and SMB display 3400–1700 Ma and 2700–650 Ma, respectively (e.g. Ghosh et al. 2004; Collins et al. 2007b, 2015; Kooijman et al. 2011; Teale et al. 2011; Plavsa et al. 2014). The ACSZ contains the detrital zircon ages from ca. 2000 to 650 Ma (Collins et al. 2007b; Sato et al. 2010) and those of TB are variable from 3000 to 1700 Ma. However, detrital zircon ages from 2000 to 1000 Ma (Collins et al. 2015) and detrital monazite ages (CHIME method) from 1900–700 Ma (Santosh et al. 2003b) indicated that the deposition of sediment has taken place up to the Neoproterozoic. The NB records detrital zircon ages from 2500–2200 Ma (Collins et al. 2007b).

As discussed in a previous section, most of the detrital zircon ages of the HC metasediments are distributed in a range of 3400–1900 Ma (Supplementary Figure 5). Thus, these Mesoproterozoic to Palaeoproterozoic protolith ages are well correlated with protolith ages of NMB and TB. The detrital ages of ca. 2800–720 Ma reported from this study and 2400–830 Ma by Sajeev et al. (2010) also well correlate with those reported in the TB since both terrains provide geochronological evidence for incorporation of Mesoproterozoic to Neoproterozoic protolith sediments.

**Correlations of metamorphic ages**

The Salem Block has subjected to high pressure granulite-facies metamorphism ($P$–$T = 14–16$ kbar and $\sim 820–860$°C) during 2500–2450 Ma (Peucat et al. 1993; Clark et al. 2009b; Anderson et al. 2012). Subsequently, Salem Block has been undergone to the middle Neoproterozoic (ca. 730 Ma) metamorphism (Bhaskar Rao et al. 1996; Ghosh et al. 2004; Collins et al. 2015). Although some workers (e.g. Ghosh et al. 2004) suggested the area has undergone to the Ediacaran–Cambrian high-grade metamorphism, the extent and $P$–$T$ evolution are poorly constrained.

The NMB has undergone to HT/UHT granulite-facies metamorphism (7–11 kbar, 950–1150°C) during 590–500 Ma (e.g. Ghosh et al. 2004; Brandt et al. 2011; Kooijman et al. 2011). The SMB has been subjected to HT/UHT metamorphism at $P = 7–11$ kbar and $T = 950–1150$°C during 550–520 Ma (Prakash et al. 2006, 2010; Collins et al. 2007b). The metamorphism of the ACSZ has been taken place at 600–500 Ma (e.g Ghosh et al. 2004; Santosh et al. 2006; Sato et al. 2010; Collin et al. 2015; Plavsa et al. 2015). The ages of HT/UHT metamorphism of the TB and the NB range from 520 to 580 Ma (Ghosh et al. 2004; Shabeer et al. 2005; Collins et al. 2007b) and 640–520 Ma (Collins et al. 2007b; Johnson et al. 2015), respectively. Hence, the metamorphic ages of ca. 665–510 Ma of the HC of Sri Lanka and the SGT of India indicate that the both Sri Lanka and SGT have undergone late Neoproterozoic metamorphism simultaneous to assembly of Gondwana.

The presence of 2000–1200 Ma Nd model ages (in the south eastern part of the SMB and the eastern margin of the ACSZ), 2800–650 Ma detrital zircon and monazite ages, 1000–750 Ma magmatic zircon ages and 600–500 Ma. Metamorphic ages in the ACSZ and SMB of the SGT and the WC of Sri Lanka may also indicate additional evidence for the Gondwana linkage. Hence, based on above petrological and geochronological relations, the possible Gondwana linkage between Sri Lanka and SGT of India is illustrated in Figure 10 and Supplementary Figure 6.

As shown in the Figure 9, the correlation of crust formation ages, magmatic ages, detrital zircon/monazite ages and metamorphic ages of the TB of India and the HC Sri Lanka clearly set the baseline for correlation of the Gondwana linkage between Sri Lanka and Southern Granulite Terrain of India.

On the other hand, several workers have considered that the Lützow–Holm Complex (LHC) of East Antarctica as an extension of the HC of Sri Lanka (e.g. Yoshida 1988; Yoshida et al. 1992; Shiraishi et al. 1994; Kriegsman 1995; Braun and Kriegsman 2003; Tsunogae et al. 2015).

Although detrital zircon ages ca. 1000 Ma have been reported from the LHC (Shiraishi et al. 1994, 2008), middle to late Neoproterozoic detrital zircon ages have not been reported so far. However, based on Sr, C, and O isotope studies, Satish-Kumar et al. (2008) suggested an
apparent age of ca. 730–830 Ma for deposition of protoliths of carbonate rocks (marbles) in the LHC. Hence, the deposition of sediments of the LHC could have taken place during the Neoproterozoic coeval to that of the HC probably providing evidence for a Neoproterozoic suture/orogenic belt extending from TB of India to LHC of Antarctica across the HC of Sri Lanka.

Conclusions

The following conclusions are reached from the present study.

1. The incorporation of ca. 2800–700 Ma detrital zircons in the studied UHT pelitic rocks indicates that the metasediments of the Highland Complex, Sri Lanka have been derived from Neoarchaean to Neoproterozoic multiple provenances. The studied metasedimentary rocks have undergone multiple thermal events from the late Cryogenian to early Cambrian (ca. 665–500 Ma) in which ca. 665 Ma and ca. 630 Ma could represent evidence for prograde metamorphism while the time span of peak UHT metamorphism is from ca. 580–530 Ma.

2. The deposition of sediment in the Highland Complex has taken place at the Neoproterozoic, atop then existing Palaeoproterozoic igneous basement of 2000–1800 Ma age and subjected to intense deformation associated with multiple thermal events during the late Cryogenian to early Cambrian (ca. 665–500 Ma) coeval with the assembly of Gondwana Supercontinent. During these events, the Palaeoproterozoic basement was reworked repetitively forming supra-crustal sequences of multiple ages in the Highland Complex.

3. The presence of similar crust formation ages, magmatic ages, detrital zircon/monazite ages and metamorphic ages in both Sri Lankan Highland Complex and the Trivandrum Block of India clearly set the baseline for correlation of the Gondwana linkage between Sri Lanka and Southern Granulite Terrain of India.

Acknowledgement

PLD appreciates the help of George Matthews, Ishwar Kumar, Vinod Samuel, and Anil Kaushik at the Indian Institute of Science, Bangalore, in zircon separations and EPMA analysis.
Disclosure statement

No potential conflict of interest was reported by the authors.

Funding

Authors are grateful to the Ministry of Technology and Research, Sri Lanka [Grant number MTR/TRD/AGR/3/1/4] and Department of Science and Technology, India for funding [Grant number DST/INT/SL/P-004]. National Research Council, Sri Lanka, [Grant number NRC-15-089] is gratefully acknowledged for extending support.

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