1 Petrogenesis of ultramafics in the Neoarchean Veligallu greenstone terrane, eastern
2 Dharwar craton, India: Constraints from bulk-rock geochemistry and Lu-Hf isotopes
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Abstract

We report on the bulk-rock geochemistry and Lu-Hf isotope systematics, and infer the petrogenesis of the ultramafic rocks in the Neoarchean Veligallu greenstone terrane, eastern Dharwar craton, India. The ultramafics along with the basalts and adakites yield an age of 2.696 ± 0.054 Ga, consistent with the available SIMS zircon U-Pb age of 2.697 ± 5 Ma, reported for the felsic volcanics, in the literature. The rocks have positive initial $\epsilon_{\text{Hf}}(2.696 \text{ Ga}) = +3.0$ to $+6.5$, consistent with an origin from a long term depleted mantle source relative to a chondritic reservoir at ~ 2.7 Ga. Geochemically, the ultramafic rocks are characterized by high $\text{MgO} = 23 - 33$ wt. %, $\text{Mg\#} \sim 89$, $\text{Cr} = 2290 - 3855$ ppm and $\text{Ni} = 604 - 966$ ppm contents; moderate to low $\text{SiO}_2 = 52 - 49$ wt. %, $\text{Al}_2\text{O}_3 = 9.2 - 4.7$ wt. %, and $\text{TiO}_2 = 0.30 - 0.14$ wt. %. To first order, these chemical compositions broadly resemble komatiites. The rocks exhibit higher than chondrite $\text{Al}_2\text{O}_3/\text{TiO}_2$ (21 – 37) ratio, depletions in their middle-REE relative to the heavy-REE i.e. $\text{Gd}_\text{N}/\text{Yb}_\text{N} < 1$, and negative Nb, Ti, and positive Zr (Hf) anomalies in a primitive mantle normalized trace element variation diagram. Hence, the Veligallu ultramafics lack the key geochemical characteristics that would indicate a plume origin typically attributed to komatiites. Post-magmatic low grade metamorphism, hydrothermal alteration and element mobility, and /or contamination by Archean upper continental crust cannot be the cause of these geochemical patterns. Further, the calculated liquid evolution lines involving major element chemistry, are inconsistent with a boninitic lineage. Present study suggests that the Veligallu ultramafic rocks may represent melts that mainly evolved through fractional crystallization of orthopyroxene $\pm$ olivine. The spatial and temporal association of these ultramafics with the arc basalts, andesites and adakites in the Veligallu greenstone terrane significantly contributes to the

1 growing evidence on the occurrence of a new class of unusual high-Mg magmatic rocks
2 presumably generated in Archean arc settings.

3 **Keywords:**

4 Ultramafic rocks

5 Lutetium-Hafnium

6 Neoarchean

7 Veligallu greenstone terrane

8 Dharwar craton

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1. Introduction

The high-Mg volcanic rocks in the Archean greenstone terranes, which are presumably generated in response to subduction-related arc magmatic processes are either boninitic in nature (e.g. Boily and Dion, 2002; Polat et al., 2002; Smithies et al., 2005; Manikyamba et al., 2005) or picritic (e.g. Polat and Hofmann, 2003). In general, boninites have high-Mg, low-TiO₂ compositions (SiO₂ > 53 wt. %, MgO > 8 wt. %, Mg# > 0.6, TiO₂ < 0.5 wt. %; Crawford et al., 1989) with higher than chondritic Al₂O₃/TiO₂ (> 22; Hickey and Frey, 1982) ratio, and occur in fore-arc (e.g. Stern and Bloomer, 1992; Ishizuka et al., 2006), back-arc (e.g. Crawford and Keays, 1987; Escrig et al., 2012) and in intra-cratonic (e.g. Srivastava et al., 2004) settings. In contrast, the high-Mg volcanic rocks that occur in association with oceanic plateau sequences are the komatiites (e.g. Puchtel et al., 1999), which are characterized by contrasting geochemical patterns, for example, sub chondritic Al₂O₃/TiO₂ (< 22), and higher Nb/Th (> 8) ratios (e.g. Chavagnac, 2004) relative to the primitive mantle. Ultramafic rocks with komatiite-like high-Mg (> 18 wt. %) compositions and boninite-type geochemical attributes, are however, extremely rare in the Archean greenstone terranes (e.g. Kerrich et al., 1998; Wilson, 2002, 2003; Szilas et al., 2012). In this contribution, we present an account of such an unusual occurrence of ultramafic rocks from the Neoarchean Veligallu greenstone terrane, eastern Dharwar craton, India.

Recently, Khanna et al. (2015) presented a detailed account of the geology, petrography and geochemistry of the metavolcanic rocks comprising of basalt, high-Mg andesite and adakite in the Veligallu terrane. The authors have shown that these volcanic rocks were generated in response to subduction-related arc magmatic processes presumably in a Neoarchean back-arc setting. The focus of this study is to present new bulk-rock Lu-Hf

isotopes and geochemical observations for the ultramafic-mafic volcanic suite in the Veligallu greenstone terrane. This work infers that : (1) the immobile trace element geochemistry indicates that these ultramafic rocks were not affected by post-magmatic alteration, low grade metamorphism and contamination by the Archean upper continental crust; (2) the geochemical attributes described by these rocks are inconsistent with a plume-related picritic or komatiitic origin, rather comparable to Phanerozoic boninites; (3) the Lu-Hf isotopes and incompatible element systematics exhibited by the ultramafics and the basalts provide a compelling evidence of a co-genetic relation between the two igneous suites; (4) Although, the trace element signatures are boninitic in nature, geochemical modeling involving major elements is inconsistent with a boninitic lineage for the Veligallu ultramafics.

2. Regional geology

The Dharwar proto-continent is subdivided into three distinct cratonic blocks: the western Dharwar craton, the eastern Dharwar craton, and the southern granulite terrane (Swami Nath and Ramakrishnan, 1981). The western and eastern sectors of the Dharwar craton comprise laterally extensive and linearly arcuate Mesoarchean and Neoarchean greenstone terranes surrounded by younger granitoids. The NNW-SSE trending shear zone extending along the eastern margin of the Chitradurga greenstone belt (Naqvi and Rogers, 1987) is taken as a marker that separates the eastern greenstone belts from those in the western sector. The study area (Fig. 1A), the Veligallu greenstone terrane, is in the eastern Dharwar craton and south of the Cuddapah Basin. The northern tip of the belt is concealed by the overlying Proterozoic sedimentary cover of the Cuddapah Basin and the southern end tapers into the surrounding granite-gneisses.

1 The Veligallu greenstone belt (Fig. 1B) broadly exhibits N-S trend with an approximate
2 strike length of ~60 km, and a maximum width of ~6 km in the central part (Srinivasan,
3 1990). The belt was subjected to regional folding and brittle-ductile deformation. Three
4 episodes of folding were recorded (Ramam and Murty, 1997). The first generation, tight
5 isoclinal steeply vertical F1 folds are represented on a mesoscopic scale in the various litho
6 units of the belt. The regional outcrop pattern, and the disposition of the various litho units
7 in the belt is attributed to the second generation F2 folds. The third generation F3 folds are
8 manifested in the form of broad warps covering the entire length of the schist belt. The
9 volcanic rocks experienced greenschist to lower amphibolite grade metamorphism. The
10 metamorphism of the volcanic units is presumed to be synchronous with the first generation
11 (F1) folds preserved in the rocks. The term 'meta' is implicit to the schist belt lithologies
12 reported in this study. The belt is surrounded by granitic gneisses and intruded by younger
13 mafic dikes and granitoids. The synkinematic emplacement of the granitoids within the
14 schist belt presumably resulted during the development of second generation F2 folds,
15 which consequently elevated the grade of metamorphism to amphibolite facies (Ramam
16 and Murty, 1997). On the contrary, the mafic magmatic intrusions do not record any such
17 deformation or metamorphism. Compositionally, these mafic intrusions vary from
18 porphyritic subalkaline basaltic dikes, pyroxenite and gabbro. The focus of this study is the
19 ultramafic igneous rocks (Fig. 1B and 2A) that occur as discontinuously interlayered
20 conformable bands paralleling the mafic volcanic rocks in the Veligallu schist belt. The
21 metavolcanic lithologies in the Veligallu terrane constitute of basalt, high Mg-andesite,
22 adakite suite of rocks that are spatially associated with banded iron formations (BIFs) and
23 subordinate metasediments. The felsic volcanic rocks, sampled elsewhere in this belt,

yielded a SIMS U-Pb zircon age of $\sim 2.697 \pm 5$ Ma (Jayananda et al., 2013), which endorses a Neoproterozoic age for the volcanic rocks in the Veligallu greenstone terrane.

3. Sampling and analytical methods

The NNE-SSW trending ultramafic rocks are best exposed in the middle part of the Veligallu schist belt. They mostly occur as low lying outcrops conformably bounded by basalts on either side and intruded by later mafic dikes. They are extremely fine grained, schistose and greenish in appearance. Nineteen samples were collected from relatively fresh portions of the outcrop devoid of quartz veins and secondary mineralization. The samples were collected from three locations from the west (N $14^{\circ} 4' 1.2''$; E $78^{\circ} 24' 46.2''$), central-west (N $14^{\circ} 2' 51.4''$; E $78^{\circ} 24' 26.12''$) and in the central (N $14^{\circ} 2' 48.7''$; E $78^{\circ} 26' 15.1''$) part. Petrographic screening was performed to ensure that the samples were free from secondary carbonate and sulphide mineralization. Moreover, the petrographic observations, did not reveal any obvious mineralogical differences between the samples collected from these three locations. The rocks were further processed for geochemistry and isotopic studies.

For bulk-rock geochemistry and Lu-Hf isotopes, the rocks were powdered manually using an agate mortar and pestle. Ten major element oxides, Cr, and Ni were analyzed using pressed powder pellets, on a Philips MagiX PRO PW2440; microprocessor controlled, wavelength dispersive sequential XRF (Table 1). The relative standard deviation for the major element oxides is $< 3\%$ (Krishna et al., 2007). Trace elements (Table 1) including large ion lithophile elements (LILE), high field strength elements (HFSE) and rare earth elements (REE) were determined by high resolution inductively coupled plasma mass spectrometry (HR-ICP-MS; Nu Instruments Attom[®], UK). The

1 detailed procedures relating to sample dissolution, analytical methodology and instrument
2 parameters are described in Khanna et al. (2015, 2016). In brief, 50 mg of finely ground
3 sample powder was digested in a freshly prepared mixture of ultrapure grade acids (HF
4 +HNO₃) taken in 3:1 ratio in screw top Teflon "Savillex" vessels, and heated on a hot plate
5 at 160°C. Certified reference material UB-N, was dissolved simultaneously following the
6 method described above and analyzed along with the samples. Oxide and oxy-hydroxide
7 ratios were low (< 0.2%) and the double charges ions ratio was < 3%. Mass bias
8 fractionation and isobaric interferences were addressed by using certified geochemical
9 reference materials. External drift correction was performed by repeated analyses of a
10 1:5000 solution of BHVO-1. Precision and accuracy are better than RSD 3% for the
11 majority of trace elements (Table A1, in appendix). The values determined for UB-N and
12 BHVO-1 are in concurrence with the recommended values of Govindaraju (1994).

13 The selected trace elements are normalized to chondrite and primitive mantle using the
14 values of Sun and Mc Donough (1989), and indicated by (N) and (pm) subscripts,
15 respectively. The major element oxides were recalculated to 100% anhydrous for inter-
16 comparison purpose. Mg# is calculated as the mole ratio of Mg/ (Mg + Fe²⁺), where Fe²⁺
17 is assumed to be 90% for the total Fe.

18 Lutetium – Hf isotopic determinations were performed on a subset of 5 ultramafics,
19 10 basalts and 3 adakites (Table 2) at the Center for Elemental Mass Spectrometry (CEMS),
20 Department of Earth and Ocean Sciences, University of South Carolina. About 100–150
21 mg of sample powders were digested in steel jacketed Parr Teflon bombs using HF–HNO₃
22 mixtures. After digestion the samples were dried three times with HNO₃ to ensure
23 digestion. The samples were then picked up in 10 ml of 1 M HNO₃ to ensure a visibly clear

1 solution. 10-15% of that solution was aliquoted quantitatively in a separate Teflon beaker
2 and spiked with mixed ^{176}Lu - ^{179}Hf and ^{149}Sm - ^{145}Nd tracers (Khanna et. al., 2014) for
3 parent-daughter determination by isotope dilution. The solution was dried down two times
4 to ensure sample-spike equilibration. The two aliquots containing the bulk sample for
5 isotope composition and isotope dilution determinations were processed identically. The
6 sample was first processed through Teflon micro columns filled with TRU spec resin
7 (Eichron). The Hf and Lu-rich cut was eluted first through the columns in 1M HNO_3 and a
8 LREE – rich fraction with 1 N HCl. Hafnium (and Lu for the isotope dilution
9 determination) was separated from that cut following the Münker et al. (2001) technique
10 using a one-step Ln-spec resin column procedure. The Sm-Nd cut was saved and results of
11 that will be reported as part of a larger isotope study.

12 Hafnium isotopes as well as Hf and Lu isotope dilution ratio measurements were
13 determined on the Thermo Finnigan NEPTUNE MC-ICP-MS with the PLUS upgrade
14 installed. The sample solutions were introduced with a 100 ul self-aspirating Teflon
15 nebulizer (ESI, USA) coupled to an APEX-Q (ESI, USA) system using a Jet-sampler cone
16 and X-skimmer cone configuration. Instrumental mass fractionation was corrected using
17 $^{179}\text{Hf}/^{177}\text{Hf} = 0.7325$. During the course of the measurements, the JMC-475 standard was
18 determined at $^{176}\text{Hf}/^{177}\text{Hf} = 0.282144 \pm 0.000008$ (2 standard deviations, $n = 10$). The
19 masses of ^{172}Yb and ^{175}Lu were monitored and the ^{176}Hf signal was corrected online using
20 the instrument software, but the corrections were negligible. Full procedural blanks were Hf
21 < 100 pg, and generally insignificant relative to the amount of Hf recovered from the
22 samples.

For $^{176}\text{Lu}/^{175}\text{Lu}$ determination we followed the method outlined by Vervoort et al. (2004), by testing the reproducibility of the natural $^{176}\text{Lu}/^{175}\text{Lu}$ ratio (0.02655) in different Lu/Yb mixtures. We determined $^{176}\text{Lu}/^{175}\text{Lu} = 0.026556 \pm 0.03\%$ ($n = 6$) ranging from Lu/Yb = 5 to 1 (concentration ratios). For fractionation correction we used $^{173}\text{Yb}/^{171}\text{Yb} = 1.129197$ (Vervoort et al., 2004; and we calculated the ^{176}Yb contribution to ^{176}Lu signal using $^{176}\text{Yb}/^{173}\text{Yb}$ (natural) = 0.793045. On the basis of a plot of ^{176}Yb -interference corrected and fractionation corrected $^{176}\text{Lu}/^{177}\text{Lu}$ vs. Yb/Lu, we determined an instrumental mass bias factor of 0.99965 on the $^{176}\text{Yb}/^{173}\text{Yb}$ ratio where the fractionation corrected $^{176}\text{Lu}/^{177}\text{Lu}$ ratios remain constant for the range of added Yb (i.e. Vervoort et al., 2004)..

We validate our complete dissolution, splitting and spiking method using multiple digestions of the USGS standard BHVO-2. The measured Hf isotope ratios and of the USGS reference material BHVO-2, is reported in Table 2, and it is identical to the literature values (e.g. Weis et al., 2005, 2007; Salters et al., 2011). The measured $^{176}\text{Lu}/^{177}\text{Hf}$ ratios for the USGS standard, BHVO-2 is given in Table 2. The reproducibility of the $^{176}\text{Lu}/^{177}\text{Hf}$ ratios is 0.3% (2 standard deviations) and the measured average BHVO-2 $^{176}\text{Lu}/^{177}\text{Hf} = 0.00880$ is well within the uncertainty of the $^{176}\text{Lu}/^{177}\text{Hf} = 0.00892$ calculated from the recommended Hf and Lu concentration in the GEOREM database. Isochrons were calculated with the ISOPLOT program (Ludwig, 2001), using the decay constants $\lambda^{176}\text{Lu} = 1.867 \times 10^{-11}$. For the isochron and initial isotope composition determinations, we assign a 0.5% error on the measured $^{176}\text{Lu}/^{177}\text{Hf}$ ratios. The initial isotope compositions were calculated using the chondritic values: $^{176}\text{Hf}/^{177}\text{Hf} = 0.282785$, $^{176}\text{Lu}/^{177}\text{Hf} = 0.0336$ (Bouvier et al., 2008).

4. Results

1 *4.1. Petrography*

2 The Veligallu ultramafic rocks predominantly consist of talc and amphibole with minor
 3 opaque oxides. Talc is flaky in nature, often occurs as subhedral pale greenish pink clusters
 4 (Fig. 2B). It characteristically exhibits one set of cleavage and straight extinction. The
 5 amphibole is predominantly anthophyllite, whereas tremolite and actinolite are present in
 6 subordinate amounts. Anthophyllite occurs as tabular grains that exhibit high relief and
 7 gray order interference colors and straight extinction (Fig. 2C). Parallel to sub parallel
 8 alignment of mainly talc and anthophyllite define the foliation in the rock. Titanite and
 9 rutile are the accessory phases that occur as disseminated minute grains in the groundmass.

10 *4.2. Bulk-rock geochemistry*

11 Collectively, the ultramafic rocks consist of moderate TiO_2 (0.14 – 0.30 wt. %), Al_2O_3
 12 (4.7 – 9.2 wt. %), Fe_2O_3 (7.1 – 9.1 wt. %); high SiO_2 (49 – 52 wt. %), MgO (24 – 34 wt.
 13 %) and $\text{Mg\#}$ (~ 89) (Table 1). Trace element geochemistry indicates high concentrations
 14 of compatible elements Cr (2290 - 3855 ppm) and Ni (634 – 966 ppm). The rocks exhibit
 15 systematic geochemical correlations between the major, trace and rare earth elements
 16 (Table A2, in appendix). In detail, the major element oxides (Al_2O_3 , TiO_2 , CaO , Na_2O and
 17 K_2O), and V, Y and Yb concentrations increase with decreasing MgO , while the more
 18 compatible Cr decreases with decreasing MgO concentrations. The rocks are characterized
 19 by high $\text{Al}_2\text{O}_3/\text{TiO}_2$ (21 – 37) ratio relative to the chondritic value (~ 22 ; Sun and
 20 McDonough, 1989), and sub chondritic Zr/Y (≤ 3). Their chondrite normalized rare earth
 21 element patterns (Fig. 3A) have elevated light-REE over the middle-REE ($\text{La}_\text{N}/\text{Sm}_\text{N} = 1.2$
 22 – 2.0), slightly negative Eu anomalies, and subtle depletions in the middle-REE relative to
 23 the heavy-REE ($\text{Gd}_\text{N}/\text{Yb}_\text{N} = 0.85 – 1.00$) imparting concave-up REE patterns. In a primitive

mantle-normalized trace element variation diagram (Fig. 3B), the samples display prominent Nb depletions with $(\text{Nb/La})_{\text{pm}}$ and $(\text{Nb/Th})_{\text{pm}} < 1$; negative Ti i.e. $(\text{Ti/Gd})_{\text{pm}} = 0.56 - 0.92$, and generally positive Zr i.e. $(\text{Zr/Sm})_{\text{pm}} = 1.0 - 1.3$, anomalies with near chondritic $\text{Zr/Hf} \sim 36$.

4.3. Lutetium-Hafnium ages of Veligallu basalt-ultramafic suite

The Lu-Hf isochron for the ten basalt samples (Fig. 4A), yields an age of 2.640 ± 0.150 Ga with an initial $^{176}\text{Hf}/^{177}\text{Hf} = 0.28124 \pm 0.00013$ ($\epsilon\text{Hf}_{(2696 \text{ Ma})} = +3.0$ to $+6.5$; Table 2).

The isochron for the five Veligallu ultramafic samples (Fig. 4B), yields an age of 2.605 ± 0.990 Ga with an initial $^{176}\text{Hf}/^{177}\text{Hf} = 0.28124 \pm 0.00065$ ($\epsilon\text{Hf}_{(2696 \text{ Ma})} = +3.2$ to $+5.6$; Table 2).

The large error in the isochron age of the ultramafics compared to the basalts (Fig. 4A), is due to the very small range in their Lu-Hf concentrations and relatively limited spread in their isotope ratios. The initial Hf isotopic composition i.e. $^{176}\text{Hf}/^{177}\text{Hf} = 0.28124$, however, is remarkably identical to that intercepted by the isochron derived for the basalt samples (Fig. 4A). The isochron for the combined basalt-ultramafic suite (Fig. 4C) gives an age of 2.679 ± 0.120 Ga with an initial $^{176}\text{Hf}/^{177}\text{Hf} = 0.281203 \pm 0.000094$. The large spread in the Lu/Hf ratios and corresponding large range in measured isotope compositions (e.g., $\epsilon\text{Hf}_{(\text{present day})} = -45$ to 167) may obscure subtle differences in eruption ages and/or initial isotope composition for the different Veligallu rock types and their sources. Given the small spread in the Lu-Hf ratio, and limited number of samples, a standalone isochron for the adakites is not conceivable. Once the adakites are regressed along with the basalt-ultramafic suite, the calculated Lu-Hf isochron for the entire group gives an age of 2696 ± 0.054 Ga with an initial $^{176}\text{Hf}/^{177}\text{Hf} = 0.281189 \pm 0.000038$ and calculated initial $\epsilon\text{Hf}_{(2696 \text{ Ma})} = 4.9 \pm 1.6$. This bulk-rock Lu-Hf age is comparatively identical to the SIMS zircon U-

Pb age of 2697 ± 5 Ma determined by Jayananda et al. (2013) for the felsic volcanic from elsewhere in the Veligallu greenstone belt. If we take into account of the isochron ages derived for the individual rock types and calculate the initials, then we get a range of initial Hf isotope ratios for all these samples. However, the initials do not change substantially, in the sense that the initial ϵ_{Hf} calculated at 2605 Ma = 4.9 ± 2.04 , is within error similar to that calculated for 2640 Ma = 4.9 ± 2.0 , and 2679 Ma = 4.9 ± 1.7 . This implies that the initial ϵ_{Hf} on an average has 0.44 ϵ_{Hf} units difference, and the isochrons have within error same initial ϵ_{Hf} isotope value that essentially requires a depleted reservoir, regardless of the ages derived for the individual rock type.

5. Discussion

5.1. Mineralogical alteration

In hydrothermally altered ultramafic igneous rocks, talc is a common alteration phase (e.g. Pawley and Wood, 1995). Addition of silica and removal of magnesium in conjunction with CO_2 addition during serpentinization can also result in the formation of talc (e.g. Naldrett, 1966). In such a case, talc often coexists with carbonate minerals e.g. magnesite or dolomite or calcite. The petrographic observations, however, do not indicate the coexistence of carbonate minerals in the Veligallu ultramafics. Furthermore, the replacement of olivine + orthopyroxene with anthophyllite and talc is consistent with a metasomatic origin at lower temperatures (< 700 °C; Abbott and Raymond, 1984). As such, the absence of relict textures i.e. orthopyroxene rimmed with anthophyllite, and olivine with serpentine, along with the observed mineral assemblage (talc + anthophyllite + tremolite $\pm$ chlorite) in these ultramafic rocks, is inconsistent with retrograde hydration of the precursor rock in response to greenschist to low grade amphibolite facie conditions of

1 regional metamorphism (Scotford and Williams, 1983). Thus, we presume that the
2 precursor rocks were subjected to purely hydrothermal metasomatic alteration.

3 *5.2. Assessing element mobility*

4 Considering that the rocks were subjected to variable degree of hydrothermal alteration
5 and low grade metamorphism, an assessment of element mobility is of critical concern
6 prior to the evaluation and petrogenetic interpretation of the rocks that occur in the Archean
7 greenstone terranes. As such, numerous early (e.g. Ludden et al., 1982), and recent studies
8 (e.g. Polat and Hofmann, 2003) have postulated that trace elements (e.g. Al, Ti, Cr, Ni, Sc,
9 V, Nb, Ta, Zr, Hf, Th, Y, and the heavy-REE) are suitable for petrogenetic interpretation
10 as they are least susceptible to mobility during hydrothermal alteration and metamorphism
11 up to amphibolite facies.

12 Although the mineralogical alterations are apparent in the sampled ultramafic rocks, the
13 robust geochemical correlation ($r > 0.75$; cf. Polat et al., 2002) between MgO and the
14 alteration sensitive Na₂O ($r = -0.94$; Table A2, in appendix), K₂O ($r = -0.87$) and CaO ($r =$
15 -0.91); and alteration resistant TiO₂ ($r = -0.84$), Al₂O₃ ($r = -0.88$), Cr ($r = 0.92$) and Yb, Y
16 ($r = -0.84$) indicates that these elements were neither significantly mobilized, nor that there
17 was any intense gain or loss of MgO during post-magmatic alteration processes. Therefore,
18 the geochemical correlations (Table A2, in appendix) in concurrence with uniform REE
19 and trace element patterns (Fig. 3) are taken to reflect primary magmatic differentiation
20 trends. Further, it has been shown that Zr can be used as an alteration-independent proxy
21 to assess the trace element mobility in recent (e.g. Pearce and Peate, 1995) and Archean
22 metavolcanic rocks (e.g. Polat et al., 2003). As a corollary, the systematic correlation

between Zr and relatively more incompatible trace elements e.g. La ($r = 0.89$), Ce ($r = 0.92$) and Sm ($r = 0.92$; Fig. 5) suggests limited elemental post magmatic mobilization.

5.3. *Petrogenetic link between the Veligallu basalts and ultramafics*

The high field strength elements Zr, Nb, Ti and Y, and the heavy-REE Gd, Yb and Lu are insensitive to the effects of hydrothermal alteration and low grade metamorphism during post magmatic crystallization processes. Hence, the concentrations and their interelement ratios can be effectively used to constrain the primary nature of the precursor melts and inspect the petrogenetic link, if any, between the various litho units that are temporally and spatially associated in a particular terrane. For instance, in Zr versus HFSE and heavy-REE bivariate diagrams (Fig. 6), the Veligallu ultramafics and basalts display linear trends with robust correlations (r = correlation coefficient) and lie within the same elemental array. This essentially indicates that the basalts and the ultramafics consist of identical range in their incompatible element ratios, and presumably crystallized from a common parental magma. Thus, inferred to be co-genetic in nature. Besides the field observation that the basalts and the ultramafic rocks are conformably associated, further evidence in support of their co-genetic relation is derived from the identical initial Hf isotopic ratios (Fig. 4) and the positive ϵHf values for the individual rocks that span an identical range between the two groups (Table 2).

5.4. *Testing for crustal contamination*

The high MgO, Mg#, Cr, and Ni contents, very low abundance in incompatible elements that are typically enriched in the crust (K, Na, Rb, Ba), as well as low HFSE (Zr, Nb, Y) and REE concentrations in these ultramafic rocks are inconsistent with modification by assimilation of crustal rocks in shallow magmatic plumbing systems. Contamination of

mafic magma by assimilation of crustal material en route to the surface will result in significantly lower Nb/Th ratio (Jochum et al., 1991) in the erupted melt compared to the primary mantle derived melt. As there is evidence for the presence of Mesoarchean upper crust in the eastern Dharwar craton (Maibam et al., 2011), to first order, the negative Nb anomalies relative to the REE i.e. $(\text{Nb/La})_{\text{pm}} < 1$ in the Veligallu ultramafics, may then reflect a crustal contamination signature. The rocks, however, do not plot on a mixing line (Langmuir et al, 1978) between the average N-MORB and the Mesoarchean continental crust (Fig. 7A), as would be expected in case of crustal contamination. Further, the initial Hf isotope compositions of the Veligallu samples do not show any systematic correlations with decreasing Nb/Th ratios (Fig. 7B) as crustal contamination would produce. The low Nb/Th ratios and negative Nb anomaly in mafic rocks is widely accepted to indicate magmatism at convergent margins in subduction-related oceanic and continental arc settings (Pearce and Peate, 1995). Accordingly, we conclude that the geochemical attributes of the Veligallu basalts and the ultramafics are inconsistent with any interaction with the Archean upper continental crust locally preserved in the Dharwar craton. The robust and concordant Lu-Hf ages defined by this diverse volcanic suite suggests that the Lu/Hf system was relatively homogenous and not significantly affected by post-magmatic deformation and metamorphism. Moreover, the ultramafic rocks are temporally and spatially associated with the basalts, high-Mg andesites and adakites in the same volcanic terrane (Khanna et al., 2015), which essentially suggests a petrogenetic link between these rocks that were presumably generated in a subduction-related oceanic arc setting.

5.5. Nature of the mantle source

Compared to the samples collected from the central part, those from the western part of the Veligallu schist belt are characterized by relatively low MgO, Mg#, Cr and Ni, and high TiO₂, Al₂O₃ and CaO contents (Table 1). The trace and rare earth element concentrations do not show such distinctive contrast between the samples collected from any of the three locations. In the Pearce Element Ratio diagrams (Fig. 8a-f; Pearce 1968), the Veligallu ultramafics essentially plot sub-parallel to orthopyroxene controlled fractionation trend with a consistent slope of 1.2, which suggests that these rocks fractionated the primary mineral phases (Ol : Opx) in the ratio 1.2 : 1 that corresponds to ~ 55 % olivine and ~ 45 % orthopyroxene. The Al₂O₃ (r = -0.88) and CaO (r = -0.91) concentrations in the Veligallu samples increase with decreasing MgO contents (see Table A2, in appendix), which is again predominantly consistent with orthopyroxene controlled low pressure fractionation of the melt (e.g. Hickey and Frey, 1982). Further, the undepleted nature of the heavy-REE ($Gd_N/Yb_N \leq 1$) in these ultramafics is consistent with their generation from a peridotitic mantle source in the spinel stability field.

The high field strength elements (Nb, Zr) along with REE (Gd, Yb), Y, Ti, V and Sc, are generally insoluble in aqueous fluids. During slab dehydration and wedge melting these elements are conserved in the slab, as such their concentrations in arc magmas are presumed to be primarily controlled by the mantle wedge (Pearce and Parkinson, 1993; Woodhead et al., 1993; Elliott et al., 1997). Furthermore, these elements and their inter element ratios are least sensitive to the effects of low pressure fractionation and crystallization processes and hence, their concentration in the derivative magmas can be used to constrain the nature of the mantle source. For example, the significantly low Ti/V (6.5 – 11.2) and Ti/Sc (38 – 62) ratios in these rocks, relative to the average N-MORB

(Ti/V = 29, and Ti/Sc = 190; Sun and McDonough, 1989), indicates that the mantle source of the Veligallu ultramafic rocks was comparatively depleted. Further, the positive initial ϵ_{Hf} compositions in the Veligallu ultramafics and basalts (Table 2) are consistent with their derivation from a long term depleted mantle source. The high $\text{Al}_2\text{O}_3/\text{TiO}_2$ (> 22) ratio in conjunction with the extremely low TiO_2 (< 0.3 wt. %) contents in the ultramafic samples is consistent with a strongly refractory nature of their mantle source (Hickey and Frey, 1982; Duncan and Green, 1987). Such depleted and refractory nature of the source can be ascribed to prior melt extraction events, perhaps at mid-ocean ridges or in back-arc basins (e.g. Falloon and Crawford, 1991; Stern et al., 1995).

5.6. *Hydrous melting of the mantle wedge*

The total REE content in the Veligallu ultramafics is significantly low ($\sum \text{REE} \sim 15$ ppm) compared to that in the basalts ($\sum \text{REE} \sim 44$ ppm; Khanna et al., 2015) and adakites ($\sum \text{REE} \sim 72$ ppm; Khanna et al., 2015). Therefore, any contribution from an extraneous source e.g. input from continental crust or subduction derived fluid/melt would significantly overprint the incompatible trace element signature of the primary melt and hence the apparent REE patterns would be essentially representative of the melt-modified mantle component. As discussed above (Section 5.4), any significant input from preexisting crustal sources is unlikely. Instead we presume that a subduction-related melt enrichment process may have essentially modified the mantle component of the Veligallu ultramafics.

In subduction-related magmatic systems, where fluid mediated melting of the mantle wedge is involved, non-conservative element transport from slab to the mantle wedge is an essential process (e.g. Elliott et al., 1997). In Th-Nb-Yb coordinate space, where Yb and

Nb are conservative relative to Th, which is extremely non-conservative (Pearce and Peate, 1995), any measureable contribution from the slab will thus result in higher Th/Nb or Th/Yb ratios relative to average N-MORB and consequently displace the sample array upwards and subparallel to the terrestrial MORB array. By analogy, the high Th/Nb, light-REE/Nb and Th/Yb (Fig. 9A), and low Nb/Th (Fig. 7B) ratios in the Veligallu ultramafics reflect the contribution from an aqueous adakitic fluid/melt derived from the slab, which presumably metasomatized the mantle wedge shortly prior to or during its melting. Similarly, the relationship between Nb/La and Hf/Sm ratios in the Veligallu samples (Fig. 9B) is consistent with their extraction from a hydrated mantle source in a subduction zone environment (e.g. Yogodzinski et al., 1995; Tate and Clarke, 1995). Thus, the Veligallu ultramafic rocks are inferred to be the products of hydrous melting of a prior depleted and refractory mantle wedge, consistent with at least a two-stage melt extraction process (e.g. Hickey and Frey, 1982; Crawford et al., 1989; Smithies et al., 2004).

5.7. *Comparison with picrites, komatiites and boninites*

The Th-Nb-La (Fig. 10A) relationship in the Veligallu samples is consistent with their eruption in an arc environment, which is comparable to island arc picrite lavas (e.g. Eggins, 1993; Polat and Hofmann, 2003). Plume-generated Archean picrites (e.g. Kitayama and Francis, 2014), on contrary, are characterized by high Nb > 12 ppm contents, Nb/Th > 8 ratio, and positive Nb anomalies (Fig. 10A). In contrast to both arc-related and plume-generated picrites, the Veligallu ultramafics are characterized by higher than chondritic $\text{Al}_2\text{O}_3/\text{TiO}_2$ (Fig. 10B) coupled with low TiO_2 contents, and sub-chondritic Ti/Sc (Fig. 10C) ratios that straddle the Phanerozoic boninite field. Therefore, it is unlikely that these ultramafic rocks are picrites.

1 The high MgO and Mg# in combination with low Al₂O₃, CaO and TiO₂ concentrations,
2 and extremely low HFSE (Nb, Zr, Y) and Yb in the Veligallu ultramafics, to first order,
3 resemble komatiites (e.g. Puchtel et al., 1997, 1999). Although, the generation of Archean
4 komatiites from a depleted mantle source with positive initial Hf compositions is not an
5 uncommon phenomena (e.g. Blichert-Toft and Arndt, 1999), the Veligallu ultramafics,
6 however, in contrast to the pristine komatiites, exhibit prominent negative Nb and Ti
7 anomalies (Figs. 3B and 10A), which is typical of volcanic arc magmas (Pearce and Peate,
8 1995; Elliott et al., 1997). Moreover the komatiites, including Al-depleted and Al-
9 undepleted type (e.g. Chavagnac, 2004), are characterized by sub to near chondritic
10 Al₂O₃/TiO₂ ratio; regardless of crustally contaminated nature (e.g. Puchtel et al., 1997;
11 Hollings and Kerrich, 1999) or not (e.g. Fan and Kerrich, 1997; Puchtel et al., 1999), which
12 contrasts the higher Al₂O₃/TiO₂ of the Veligallu ultramafics (Fig. 10B).

13 Based on the geochemical similarities between the modern boninites and Barberton
14 komatiites and komatiitic basalts (e.g. the low TiO₂ < 0.5 wt. % contents and low Ti/Zr
15 ratios, the Nb and Ti depletions, the positive Zr anomaly and flat to U-shaped REE
16 patterns), Parman et al. (2001, 2004) argued for a subduction zone origin for the Barberton
17 komatiites. According to their model, the Barberton komatiites were produced by hydrous
18 melting in a mantle plume in a similar manner in which the modern boninites are generated
19 by hydrous melting of a mantle wedge in an arc setting; except that the mantle temperatures
20 were relatively higher in the Archean, and that these komatiites represent Archean
21 equivalents of modern boninites. On the basis of typical geochemical fingerprints of
22 subduction related processes, the argument, however, is contested (see Chavagnac, 2004
23 for a review).

We tested the above model in the case of the Veligallu ultramafics. On the basis of following evidence, we find that such a model is in contrast to the field observations and geochemical nature of the volcanic units in the Veligallu greenstone terrane: (1) the ultramafic rocks are spatially associated with the basalts that are geochemically identical to modern back-arc type tholeiites (Khanna et al., 2015). Collectively, the basalt - high-Mg andesite - adakitic volcanic rocks in the Veligallu terrane provide geochemical evidence for a subduction-related magmatism (Khanna et al., 2015; (2) neither the basalts nor the ultramafic rocks provide any textural evidence of a komatiitic affinity (Kerr and Arndt, 2001); (3) the Veligallu ultramafics have higher $\text{Al}_2\text{O}_3/\text{TiO}_2$ (~ 30) ratio at a mean value of 0.21 wt. % TiO_2 versus the lower $\text{Al}_2\text{O}_3/\text{TiO}_2$ (~ 10) ratio for a minimum value of 0.31 wt. % TiO_2 in the Barberton komatiites (Fig. 10B; Parman et al., 2004). The Barberton komatiites are characterized by higher Nb/Th (8.3 – 14; Chavagnac, 2004) ratio and plot within the MORB array, which does not relate to a subduction origin, and in stark contrast to that in the Veligallu samples (Nb/Th = 1.4 – 2.1); (4) the range in $\text{CaO}/\text{Al}_2\text{O}_3$ (0.4 – 1.05) and CaO/TiO_2 (14 – 38) ratios in the Veligallu samples is identical to that observed in certain high –calcium boninites ($\text{CaO}/\text{Al}_2\text{O}_3$ = 0.49 – 0.9 and CaO/TiO_2 = 17 – 48; Falloon and Crawford, 1991); moreover, (5) the low TiO_2 (< 0.5 wt. %) coupled with higher than chondritic $\text{Al}_2\text{O}_3/\text{TiO}_2$ (> 22), and sub chondritic Ti/Sc ratios in combination with negative Nb, and Ti anomalies, and reversely fractionated heavy-REE with U-shaped appearance, are comparable to boninites.

6. Petrogenetic implications

Even though, the trace element systematics of the Veligallu ultramafics are boninitic in nature, the significantly high MgO, Mg# and low SiO_2 contents, do not comply with the

proposed standard classification schemes for boninites (Le Bas 2000; Verma et al., 2002). On the contrary, the geochemical compositions in the Phanerozoic boninites seldom exceed 18 wt. % MgO (e.g. Brown and Jenner, 1989; König et al., 2010; Xia et al., 2012), the SiO₂ content in the Archean boninites, in contrast to the modern boninites, by and large is < 52 wt. % (e.g. Kerrich et al., 1998; Boily and Dion, 2002; Polat et al., 2002; Smithies et al., 2004; Manikyamba et al., 2005). A maximum MgO content of ~25 wt. % with a minimum ~44 wt. % SiO₂ has been reported for the boninitic rocks from the Neoarchean Abitibi greenstone belt (Kerrich et al., 1998). Therefore, one to one extrapolation between the major element chemistry of the Archean and the Phanerozoic compositional types must be applied cautiously. Alternately, we have examined the following petrogenetic models from the literature, to further evaluate the petrogenesis of the Veligallu ultramafics.

Recently, in the Cambrian-early Ordovician Niagara Icefalls, northern Victoria Land, Antarctica, Tribuzio et al. (2008) reported the major and trace element composition of the clinopyroxenes from the mafic-ultramafic cumulate sequence consisting of dunite, harzburgite, orthopyroxenite, melagabbonorite and gabbonorite. On the basis of textural relations and major element compositions they proposed a cumulus origin for the Niagara Icefalls mafic-ultramafic sequence. They extrapolated the incompatible element signatures and rare earth element attributes, to suggest their origin from boninite-type melts. By analogy, they concluded that primitive boninites (Bloomer and Hawkins, 1987) consist of olivine + orthopyroxene + Cr-spinel which give rise to high-Mg low-Si compositions in boninites that are comparable to the Niagara Icefalls harzburgites. In contrast, the relatively high-Si low-Mg compositional types consist of orthopyroxene ± plagioclase ± clinopyroxene can thus be correlated with the gabbonorites of their study. As such they

1 attributed the chemical variation between the major elements of the Niagara Icefalls
2 sequence to a fractional crystallization process controlled by the separation of olivine,
3 orthopyroxene, clinopyroxene, Cr-spinel and plagioclase in the melt, following the model
4 proposed for the genesis of boninite series volcanic rocks (e.g. Bloomer and Hawkins,
5 1987; Taylor et al., 1994).

6 In the ~3.7 Ga Isua supracrustals belt, southern West Greenland, Szilas et al. (2015)
7 made a comprehensive study of the basalts and boninites. The ultramafic samples have
8 compositions of 26 to 48 wt. % MgO and $\text{TiO}_2 < 0.5$ wt. %, and extend to higher than
9 chondritic $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratio (19 – 180). On the basis of MgO vs. $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratio (see Fig.
10 4 in Szilas et al., 2015 for a review), they classified these ultramafic rocks under two
11 distinct groups that plot towards tholeiitic and boninitic fractionation trends, respectively.
12 The chemical compositions of these Isua rocks are comparable to those in the Veligallu
13 ultramafics, particularly, the Isua ultramafic rocks that follow tholeiitic basalt fractionation
14 trend with high MgO ~ 33 wt. %, low Al_2O_3 (0.6 – 9.6 wt. %), $\text{TiO}_2 \sim 0.2$ wt. % coupled
15 with higher than chondritic $\text{Al}_2\text{O}_3/\text{TiO}_2 \sim 31$, ratio. By modeling the liquid evolution line
16 using the MELTS software (Ghiorso and Sack, 1995; Asimow and Ghiorso, 1998), Szilas
17 et al. (2015) suggested that the basalts and their derivative ultramafic rocks can be
18 genetically linked to fractional crystallization processes, suggestive of a cumulate origin
19 for the ultramafic rocks.

20 Similar geochemical modeling using the MELTS software (Ghiorso and Sack, 1995;
21 Asimow and Ghiorso, 1998) was performed in the case of Veligallu basalt-ultramafic co-
22 genetic suite, with the following parameters (Szilas et. al., 2014). A fractional melting
23 model from 1310 °C to 1140 °C involving 10 °C intervals was used. Oxygen fugacity was

1 buffered at QFM, at a constant pressure of 2 Kb. Initial H₂O was arbitrarily taken as 0.5
 2 wt. %. The liquid line of descent for the mean composition of the Veligallu basalts was
 3 calculated to test if the ultramafics are cumulates of the basalts. The geochemical
 4 compositions of the Veligallu basalts and the ultramafics along with the modeled
 5 corresponding liquid and cumulate compositions are shown in the Figure 11. The fractional
 6 liquid melt and the derivative bulk solid / cumulate for a Phanerozoic boninite sample
 7 (Hickey and Frey, 1982) is also modeled and shown for comparison.

8 The calculated liquid compositions were able to reproduce the observed major element
 9 variations in the Veligallu basalts. The corresponding TiO₂, Al₂O₃ and CaO contents in the
 10 derived orthopyroxene cumulate, however, appear to be even more depleted than those in
 11 the Veligallu ultramafics (Figs. 11a-c). The key ratios between the major element oxides
 12 in the modeled liquids and their derivative orthopyroxene cumulates, however, are in good
 13 agreement with those determined for the Veligallu basalts and ultramafics (Figs. 11d-f).
 14 These ratios distinguish between boninites from rest of the magmatic rocks. It is observed
 15 (Figs. 11d-f) that the Veligallu ultramafics follow tholeiitic basalt fractionation trend,
 16 therefore, may not qualify as “boninites” *sensu stricto*.

17 The bulk-rock geochemical composition of the Veligallu ultramafics is relatively fertile
 18 and in stark contrast to that essentially observed in the harzburgites and /or
 19 orthopyroxenites, which are extremely depleted in their major and trace elements (e.g.
 20 Al₂O₃ < 0.3 wt. %, TiO₂ < 0.02 wt. %, CaO < 0.07 wt. %, at ~48 – 34 wt. % MgO; Zr < 1
 21 ppm, Nb < 0.5 ppm, Y < 1 ppm, Sc = 3-11 ppm and V = 8-27 ppm; Jaques and Chappell,
 22 1980). One such explanation for the comparatively higher abundance of Al₂O₃, CaO and
 23 TiO₂ in the Veligallu samples between the observed compositions and the calculated

corresponding cumulates, could be due to the accumulation of orthopyroxene or plagioclase in the melt, or perhaps due to cumulate-melt interaction, because such depleted compositions will readily have their incompatible elements overprinted by small degrees of entrapped liquid or percolating melts that may yield anomalous concentrations in the incompatible elements (Szilas et al., 2014, 2015). To first order, it is a possibility that the percolating melts may modify the incompatible trace element signatures, but it is not clear about what kind of percolating melt could affect a change in the major element composition of a cumulate?. Further, the petrographic studies did not indicate any accumulation of crystals e.g. orthopyroxene or plagioclase, nor any entrapped melt.

Since the Veligallu ultramafics are fine grained, schistose and metamorphosed, the primary textures have been obliterated. When olivine and orthopyroxene is hydrated and serpentinized, the rock eventually becomes very fine grained. In such rocks if the exsolved chromite or magnetite is found along the crystal margins it indicates the possibility of a cumulate (e.g. Fig. 3 in Szilas et al., 2014). Although mineral chemistry indicated the presence of chrome spinel (Cr_2O_3 ~54 wt. %, FeO ~33 wt. %, Al_2O_3 ~5.3 wt. %) in the Veligallu ultramafics, such exsolution texture is not identified.

7. Conclusions

Through this study we briefly conclude that : (1) The rocks have positive initial $\epsilon_{\text{Hf}}^{(2.696 \text{ Ga})} = + 3.0$ to $+ 6.5$, consistent with an origin from a long term depleted mantle source relative to a chondritic reservoir at ~2.7 Ga; (2) the Veligallu ultramafic rocks may represent melts that mainly evolved through fractional crystallization of orthopyroxene $\pm$ olivine; (3) the trace element signatures are akin to the boninitic magmas derived from the partial melting of a refractory and hydrated peridotitic mantle wedge (Hickey and Frey,

1 1982); and (4) The spatial and temporal association of these ultramafics with the arc
2 basalts, andesites and adakites in the Veligallu greenstone terrane significantly contributes
3 to the growing evidence on the occurrence of a new class of unusual high-Mg magmatic
4 rocks presumably generated in arc settings (Wilson, 2002, 2003; Szilas et al., 2012).

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Figure captions

Fig. 1 (A) Simplified geological map of the southern peninsular India comprising of three major tectonic blocks: the western Dharwar craton (WDC), the eastern Dharwar craton (EDC) and the southern granulite terrane (SGT). Also shown in the box is the location of Veligallu greenstone belt in the eastern Dharwar craton. **(B)** Generalized geological map

of the Veligallu greenstone belt, modified after Srinivasan (1990), showing the sample locations. The sample locations shown for basalts, andesites and adakites are discussed in Khanna et al. (2015).

Fig. 2A Outcrop of Veligallu ultramafics. Photomicrographs of the Veligallu ultramafics showing **(B)** pinkish green cluster of talc, and **(C)** tabular grains of anthophyllite.

Fig. 3 (A) Chondrite normalized rare earth element, and **(B)** primitive mantle normalized trace element variation diagram for the Veligallu ultramafics.

Fig. 4 Plot of $^{176}\text{Lu}/^{177}\text{Hf}$ vs. $^{176}\text{Hf}/^{177}\text{Hf}$ bulk-rock isochrons, regressed for Veligallu arc basalt-ultramafic suite, and adakites; using the ISOPLOT program of Ludwig (2001). See text for details.

Fig. 5 Zr vs. select incompatible trace elemental behavior of the Veligallu ultramafics. See text for details.

Fig. 6 Zr versus high field strength elements, **(A)** Nb, **(B)** Ti, **(C)** Y, and the heavy-REE **(D)** Yb, bivariate diagram for the Veligallu basalt-ultramafic suite. The fractionation trends are consistent with co-genetic nature. The stippled parallel diagonal lines are arbitrary lines that represent the restricted range between the two suites. See text for details.

Fig. 7 (A) Nb vs. La/Nb bivariate diagram modified after Smithies et al. (2005); and **(B)** Nb/Th vs. initial $\epsilon\text{Hf}_{(t = 2696 \text{ Ma})}$ plot for the basalt-ultramafic suite, and adakites. The geochemical attributes of the Veligallu samples are inconsistent with assimilation of Mesoarchean upper crust present in the Dharwar craton, India (TTG; Naqvi et al., 2009). Archean upper continental crust (AUCC; Rudnick and Gao, 2004) is also shown for comparison. The data for the Mariana arc basalts is from Pearce et al. (2005) and Reagan

et al. (2010). The values for N-MORB are from Sun and McDonough (1989). See text for details.

Fig. 8 Plots of molar proportions of major, and trace element concentrations for the Veligallu basalt-ultramafic suite. See text for details.

Fig. 9 (A) Nb/Yb vs. Th/Yb ratio-ratio discrimination diagram after Pearce and Peate (1995), which distinguishes between the contribution from slab and wedge derived components. **(D)** Primitive mantle normalized Nb/La vs. Hf/Sm diagram for Veligallu ultramafics (this study), basalts and adakites (Khanna et al., 2015). Figure after Wang et al. (2015) and fields are from La Flèche et al. (1998). Fields cited from: A = Yogodzinski et al. (1995); B = Francalanci et al. (1993); C = Tate and Clarke (1995); D = Woolley et al. (1991); E = Dupuy et al. (1992), Furman (1995). The data for the Mariana arc basalts is from Pearce et al. (2005) and Reagan et al. (2010). See text for details.

Fig. 10 (A) Primitive mantle normalized Nb/Th vs. Nb/La diagram for Veligallu ultramafics (this study) and basalts (Khanna et al., 2015). Fields for island arc lavas, oceanic plateau lavas, and MORB lavas is as shown in Puchtel et al. (1999). The data envelope of Mariana arc basalts includes fore-arc (Reagan et al., 2010) and back-arc (Pearce et al., 2005) basalts. **(B)** Bivariate TiO_2 vs. $\text{Al}_2\text{O}_3/\text{TiO}_2$, and **(C)** Ti/Sc vs. Ti/V ratio-ratio, discrimination diagrams for Veligallu ultramafics. The MORB field in **(C)** is adopted from Hickey and Frey (1982). Note the contrasting geochemistry between the Archean and Phanerozoic high-Mg volcanic rocks. The data sources for Phanerozoic boninites are from: Hickey and Frey (1982); Brown and Jenner (1989); Cameron (1989); Crawford et al. (1989); Coish (1989); Rogers et al. (1989); König et al. (2008, 2010); Reagan et al. (2010). Data for Phanerozoic picrite and komatiite is from Eggins (1993) and

1 Kerr et al. (1996), respectively. Archean komatiites are from: Puchtel et al. (1999;
2 Sumozero-Kenozero) and Parman et al. (2004; Barberton). Archean picrites are from: Polat
3 and Hofmann (2003) and Kitayama and Francis (2014). The vertical and horizontal stippled
4 lines are chondrite values (Sun and McDonough, 1989). See text for details.

5 **Fig. 11** Veligallu basalt (Khanna et al., 2015) and ultramafic data from this study, plotted
6 together with MELTS calculations of the liquid evolution and the corresponding bulk
7 cumulates during fractional crystallization of orthopyroxene. Liquid evolution line for a
8 boninite sample from Hickey and Frey (1982) is also modelled and shown for comparison.
9 See text for details.

Fig.1

Fig. 2

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Fig. 10

Table 1

Major (wt. %) and trace (ppm) element composition of the Veligallu ultramafic rocks, eastern Dharwar craton, India

	Central part								Central-west						Western part		
	TVKR2	TVKR3	TVKR4	TVKR9	TVKR10	TVKR12	TVKR15	TVKR16	STV1	STV2	STV3	STV4	STV5	STV6	TVKR17	TVKR18	TVKR20
SiO ₂	51.01	51.31	50.40	51.51	50.72	49.35	50.37	49.99	51.96	49.47	50.96	50.40	49.34	49.73	49.23	49.51	48.60
TiO ₂	0.14	0.16	0.21	0.22	0.15	0.20	0.19	0.23	0.22	0.24	0.25	0.20	0.24	0.20	0.30	0.29	0.26
Al ₂ O ₃	5.30	4.95	5.70	5.23	5.35	6.36	5.31	5.85	4.83	5.77	5.16	4.72	5.43	4.84	8.22	8.50	9.19
Fe ₂ O ₃	7.30	7.42	7.45	7.24	7.27	7.82	7.17	7.13	9.07	8.87	8.79	8.53	8.36	8.46	7.84	8.20	8.18
MnO	0.11	0.11	0.13	0.11	0.12	0.11	0.11	0.14	0.12	0.15	0.14	0.13	0.15	0.16	0.14	0.14	0.14
MgO	33.74	32.41	31.29	31.35	32.54	30.43	33.05	30.36	30.71	30.24	30.34	30.61	29.51	30.74	26.72	26.16	23.55
CaO	2.29	3.55	4.64	4.23	3.78	5.55	3.72	6.03	2.99	5.12	4.16	5.27	6.73	5.72	7.16	6.85	9.66
Na ₂ O	0.04	0.04	0.08	0.06	0.03	0.09	0.03	0.15	0.05	0.08	0.08	0.07	0.13	0.08	0.25	0.22	0.26
K ₂ O	0.02	0.01	0.05	0.03	0.01	0.05	0.02	0.09	0.02	0.03	0.08	0.03	0.07	0.06	0.09	0.09	0.11
P ₂ O ₅	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.05	0.03	0.03	0.03	0.04	0.04	0.04
Mg#	90	90	90	90	90	89	90	89	88	88	88	89	88	89	88	87	87
Cr	3784	3688	3502	3855	3540	3490	3640	3280	3182	3581	3544	3204	3124	3431	2773	2969	2290
Ni	916	924	953	880	966	965	914	604	774	728	677	737	715	634	700	813	731
Rb	0.92	0.26	1.75	0.99	0.55	2.78	0.91	3.61	1.05	2.39	3.66	1.38	2.67	2.64	3.65	4.00	3.31
Sr	2.87	2.45	4.65	3.95	3.51	6.81	4.15	9.36	6.82	24.50	6.68	9.60	9.18	8.10	5.80	7.09	6.15
Cs	0.34	0.13	0.37	0.30	0.35	0.29	0.18	0.08	0.06	0.22	0.15	0.11	0.26	0.16	0.14	0.17	0.16
Ba	7	5	9	6	4	15	5	19	15	52	16	10	14	14	14	16	17
Sc	26	18	29	26	25	33	21	29	25	25	22	24	25	26	30	38	32
V	127	82	143	125	124	161	103	151	119	124	132	115	128	139	153	195	167
Nb	0.99	0.63	1.14	0.74	0.68	1.10	0.58	0.87	0.93	1.08	0.95	0.83	0.79	0.78	0.87	0.93	1.22
Zr	18	13	21	18	16	24	14	25	23	27	22	22	24	23	21	24	28
Hf	0.49	0.36	0.59	0.50	0.46	0.67	0.39	0.74	0.65	0.77	0.64	0.62	0.67	0.67	0.60	0.69	0.79
Th	0.50	0.32	0.55	0.52	0.42	0.68	0.37	0.52	0.66	0.54	0.55	0.47	0.54	0.56	0.56	0.67	0.75
U	0.23	0.11	0.19	0.22	0.31	0.28	0.13	0.19	0.21	0.21	0.19	0.21	0.19	0.20	0.31	0.28	0.36
Y	5.8	5.5	7.2	5.9	5.6	8.4	4.9	8.3	7.5	8.9	7.5	7.1	8.3	8.5	8.4	9.9	9.4
La	1.10	1.17	2.31	1.41	1.29	2.74	0.96	2.48	1.96	2.85	2.14	2.34	1.82	1.97	1.70	2.09	2.66
Ce	2.72	2.54	4.57	2.88	2.68	5.14	2.37	4.81	3.68	5.63	4.54	4.42	4.19	3.80	3.63	4.95	6.04
Pr	0.38	0.36	0.66	0.41	0.41	0.74	0.34	0.67	0.55	0.70	0.59	0.69	0.60	0.55	0.53	0.68	0.77
Nd	1.76	1.60	2.84	1.81	1.84	3.08	1.56	2.91	2.79	3.00	2.47	2.87	2.86	2.80	2.35	2.95	3.29
Sm	0.55	0.51	0.77	0.59	0.54	0.90	0.46	0.84	0.90	0.87	0.68	0.79	0.88	0.91	0.74	0.79	0.92
Eu	0.16	0.15	0.24	0.18	0.16	0.25	0.15	0.30	0.31	0.26	0.26	0.29	0.39	0.32	0.27	0.29	0.32
Gd	0.65	0.62	0.90	0.67	0.65	0.99	0.57	1.10	1.08	1.06	0.95	0.96	1.08	1.10	0.93	1.13	1.13
Tb	0.13	0.12	0.16	0.13	0.13	0.20	0.11	0.21	0.20	0.20	0.17	0.17	0.20	0.20	0.17	0.22	0.22
Dy	0.86	0.84	1.13	0.91	0.88	1.30	0.74	1.32	1.24	1.34	1.16	1.12	1.29	1.26	1.24	1.50	1.48
Ho	0.20	0.19	0.25	0.21	0.20	0.31	0.18	0.30	0.28	0.31	0.26	0.25	0.29	0.28	0.28	0.34	0.34
Er	0.58	0.52	0.73	0.59	0.55	0.83	0.48	0.86	0.79	0.87	0.75	0.72	0.83	0.80	0.80	0.99	0.93
Tm	0.09	0.08	0.12	0.10	0.09	0.13	0.08	0.14	0.13	0.13	0.12	0.11	0.13	0.14	0.13	0.16	0.15
Yb	0.63	0.57	0.78	0.63	0.60	0.91	0.53	0.96	0.89	0.93	0.80	0.79	0.92	0.94	0.89	1.07	1.02
Lu	0.10	0.09	0.13	0.11	0.10	0.15	0.08	0.15	0.14	0.14	0.13	0.13	0.15	0.15	0.15	0.17	0.17

Table 2

Bulk-rock Lu-Hf isotope data for the Veligallu basalts, adakites and ultramafic rocks.

Simultaneous determinations for the certified reference material BHVO-2 are also shown.

Rock type	Sample No.	$^{176}\text{Hf}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$	Lu (ppm)	Hf (ppm)	$\varepsilon\text{Hf}_{(t=2696\text{ Ma})}$
Ultramafic	TVKR2	0.282679 ± 07	0.0287	0.0736	0.3639	+ 5.16
	TVKR4	0.283022 ± 14	0.0356	0.0941	0.3757	+ 4.82
	TVKR12	0.282814 ± 09	0.0323	0.1263	0.5559	+ 3.48
	TVKR17	0.282989 ± 06	0.0345	0.1299	0.5345	+ 5.61
	TVKR20	0.282942 ± 10	0.0349	0.1404	0.5715	+ 3.22
Basalt *	TVK67	0.283870 ± 08	0.0516	0.3641	1.0011	+ 5.46
	TVK68	0.283435 ± 10	0.0426	0.3282	1.0929	+ 6.51
	TVK72	0.283423 ± 13	0.0431	0.3630	1.1971	+ 5.34
	TVK74	0.282865 ± 08	0.0318	0.3988	1.7800	+ 6.12
	TVK77	0.283047 ± 12	0.0357	0.3985	1.5868	+ 5.53
	TVK81	0.283192 ± 05	0.0381	0.4043	1.5052	+ 6.14
	TVK87	0.283398 ± 07	0.0420	0.4027	1.3629	+ 6.46
	TVK137	0.282580 ± 06	0.0275	0.2499	1.2893	+ 3.87
	TVK155	0.284252 ± 15	0.0604	0.3222	0.7578	+ 3.00
	TVK156	0.283328 ± 14	0.0422	0.3076	1.0343	+ 3.46
Adakite *	TVK46	0.281318 ± 08	0.0029	0.0759	3.6624	+ 4.11
	TVK47	0.281323 ± 04	0.0024	0.0629	3.7116	+ 5.10
	TVK78	0.281314 ± 05	0.0024	0.0608	3.6662	+ 5.19
BHVO-2	replicate 1	0.283098 ± 07	0.0088	0.2671	4.3057	--
	replicate 2	0.283106 ± 06	0.0088	0.2674	4.3130	--
	replicate 3	0.283101 ± 09	0.0088	0.2666	4.2958	--
	replicate 4	0.283101 ± 05	0.0088	0.2686	4.3383	--
	replicate 5	0.283101 ± 08	0.0088	0.2693	4.3547	--

* the geochemical dataset for these samples is given in Khanna et al. (2015)

Table A1

Analytical data, relative standard deviations (RSD) for UB-N and BHVO-1 obtained by HR-ICP-MS.

Analyte	Mass No.	UB-N			BHVO-1		
		A	B	% RSD	A	B	% RSD
Sc	45	12.790	13.000	1.15	31.16	31.80	2.56
V	51	74.880	75.000	0.11	307.88	317.00	2.98
Cu	63	27.427	28.000	1.45	134.34	136.00	1.55
Zn	66	84.358	85.000	0.53	99.68	105.00	3.15
Ga	71	3.038	3.000	0.89	20.81	21.00	1.44
Rb	85	3.997	4.000	0.05	10.95	11.00	0.97
Sr	88	8.893	9.000	0.85	395.07	403.00	1.58
Y	89	2.465	2.500	1.00	27.32	27.60	1.29
Zr	90	3.853	4.000	2.65	177.39	179.00	1.79
Nb	93	0.052	0.050	2.77	18.94	19.00	1.42
Cs	133	9.909	10.000	0.65	0.12	0.13	2.41
Ba	137	26.714	27.000	0.75	139.48	139.00	1.79
La	139	0.335	0.350	3.10	15.71	15.80	2.23
Ce	140	0.786	0.800	1.25	38.90	39.00	2.22
Pr	141	0.121	0.120	0.59	5.71	5.70	2.48
Nd	146	0.585	0.600	1.79	25.40	25.20	2.17
Sm	147	0.207	0.200	2.43	6.34	6.20	4.09
Eu	151	0.077	0.080	2.70	2.06	2.06	3.21
Gd	157	0.293	0.300	1.67	6.35	6.40	1.84
Tb	159	0.059	0.060	1.19	0.96	0.96	3.29
Dy	163	0.383	0.380	0.56	5.20	5.20	3.41
Ho	165	0.091	0.090	0.78	0.99	0.99	3.21
Er	166	0.287	0.280	1.75	2.40	2.40	3.10
Tm	169	0.049	0.049	0.14	0.33	0.33	2.80
Yb	172	0.274	0.280	1.53	2.04	2.02	2.94
Lu	175	0.044	0.045	1.59	0.29	0.29	3.92
Hf	178	0.097	0.100	2.15	4.43	4.38	3.10
Ta	181	0.019	0.020	1.79	1.25	1.23	2.12
Pb	208	12.877	13.000	0.67	2.51	2.60	9.28
Th	232	0.068	0.070	2.05	1.08	1.08	4.11
U	238	0.072	0.070	1.99	0.43	0.42	5.20

A – values from ICP-MS (average of 4 values)

B – values from Govindaraju (1994) and GEOREM (georem.mpch-mainz.gwdg.de)

Table A2

Corelation matrix for the Veligallu ultramafic rocks, eastern Dharwar craton, India.

	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MgO	CaO	Na ₂ O	K ₂ O	Cr	Ni	Sc	V	Nb	Zr	Hf	Th	Y	Ce	Nd	Gd	Dy	Yb
SiO ₂	1.00																					
TiO ₂	-0.52	1.00																				
Al ₂ O ₃	-0.72	0.69	1.00																			
Fe ₂ O ₃	-0.07	0.43	0.43	1.00																		
MgO	0.71	-0.84	-0.88	-0.40	1.00																	
CaO	-0.89	0.71	0.79	0.22	-0.91	1.00																
Na ₂ O	-0.78	0.84	0.91	0.17	-0.94	0.90	1.00															
K ₂ O	-0.74	0.85	0.77	0.29	-0.87	0.86	0.91	1.00														
Cr	0.70	-0.71	-0.78	-0.43	0.92	-0.86	-0.86	-0.76	1.00													
Ni	0.32	-0.58	-0.14	-0.52	0.46	-0.49	-0.48	-0.63	0.50	1.00												
Sc	-0.62	0.55	0.77	0.05	-0.67	0.62	0.72	0.67	-0.54	-0.08	1.00											
V	-0.67	0.65	0.78	0.16	-0.73	0.68	0.77	0.79	-0.58	-0.25	0.97	1.00										
Nb	-0.38	0.32	0.51	0.27	-0.51	0.41	0.47	0.56	-0.47	-0.07	0.66	0.66	1.00									
Zr	-0.61	0.63	0.51	0.55	-0.74	0.74	0.68	0.80	-0.73	-0.62	0.68	0.75	0.70	1.00								
Hf	-0.59	0.63	0.48	0.54	-0.72	0.72	0.67	0.80	-0.69	-0.66	0.67	0.75	0.67	1.00	1.00							
Th	-0.47	0.59	0.61	0.52	-0.73	0.58	0.62	0.67	-0.64	-0.31	0.77	0.80	0.81	0.87	0.85	1.00						
Y	-0.70	0.76	0.65	0.53	-0.84	0.80	0.80	0.86	-0.76	-0.59	0.77	0.84	0.60	0.92	0.92	0.83	1.00					
Ce	-0.64	0.58	0.56	0.44	-0.74	0.74	0.66	0.80	-0.68	-0.43	0.69	0.74	0.78	0.91	0.90	0.79	0.86	1.00				
Nd	-0.59	0.54	0.41	0.58	-0.66	0.68	0.57	0.68	-0.67	-0.50	0.64	0.68	0.69	0.95	0.94	0.80	0.89	0.93	1.00			
Gd	-0.54	0.64	0.40	0.67	-0.69	0.67	0.61	0.73	-0.69	-0.69	0.59	0.66	0.55	0.95	0.95	0.78	0.94	0.84	0.94	1.00		
Dy	-0.65	0.73	0.62	0.56	-0.82	0.78	0.77	0.83	-0.77	-0.58	0.75	0.81	0.63	0.95	0.95	0.84	0.99	0.89	0.93	0.96	1.00	
Yb	-0.64	0.71	0.58	0.56	-0.78	0.74	0.74	0.81	-0.74	-0.62	0.74	0.80	0.59	0.94	0.95	0.83	0.99	0.85	0.91	0.97	0.99	1.00