



Review Article

A record of Archean accretionary-to-collisional orogenesis and building of a habitable continent in the Dharwar craton, southern India

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ABSTRACT

This contribution presents a synthesis of tectonic fabrics, metamorphic, geochronologic, elemental and Nd-Hf isotope data addressing evolution of Archean crust through accretionary to collisional orogenesis in the Dharwar craton. A large crustal corridor was chosen across the Chitradurga boundary shear zone covering western and central blocks forming a wide time window into the Archean Earth.

The western block preserves 3430–3150 Ma TTG-type granitoids with 3380–3150 Ma komatiite volcanics, 3000–2850 Ma Kibbanahalli arm and 2740–2677 Ma Chitradurga belt. P-T data reveal two thermal events with a 2960 Ma event [6–8 kbar/500–690°C] and a 2620 Ma high-pressure event with 12–13 kbar/350–450°C. Elemental characteristics of TTGs with Nd-Hf isotope [$\epsilon_{\text{Nd(T)}} = +2.9$ to -0.9 ; in-situ Hf ($\epsilon_{\text{Hf(T)}} = +5.9$ to 0.0)] reveal their origin by melting of newly formed mafic arc crust. The low to moderate incompatible elements coupled with Nd isotopes [$\epsilon_{\text{Nd(T)}} = +2.7$ to -1.0] of komatiites suggest their origin from heterogeneous deeper mantle. Higher incompatible element contents (total REE = 28.27–158.41 ppm) and Nd isotopes [$\epsilon_{\text{Nd(T)}} = -0.2$ to -1.5] of Kibbanahalli arm volcanics suggest their derivation from primitive mantle. Collision of oceanic plateaus, arcs, granitoid crust caused crustal thickening, lower crustal melting causing generation of ca. 3000–2960 Ma potassic granites. Elemental and Nd isotope data [$\epsilon_{\text{Nd(T)}} = +3.9$ to -4.1] of Chitradurga volcanics point depleted to primitive mantle source with crustal contamination. Elemental and Nd isotope [$\epsilon_{\text{Nd(T)}} = +1.1$ to -7.2] data of ca. 2600 Ma potassic granites are attributed to melting of thickened crust in a collisional setting.

The central block comprises 2740 Ma Javagondanahalli greenstone belt, minor 3400–3100 Ma TTGs with widespread 2680–2600 Ma transitional TTGs and 2560–2500 Ma sanukitoids, 2500 Ma anatectic granites. The crust was affected by three metamorphic events during 3140 Ma, 2630 Ma and 2500 Ma. Elemental and Nd isotopes [$\epsilon_{\text{Nd(T)}} = +2.6$ to -0.4] of TTGs reveal their origin by melting of thickened arc crust. Incompatible elements and Nd isotope [$\epsilon_{\text{Nd(T)}} = +0.1$ to -2.3] data of Javagondanahalli volcanics attributed to their origin from depleted mantle with crustal contamination. The elemental and Nd isotope [$\epsilon_{\text{Nd(T)}} = +2.0$ to -1.6] of transitional TTGs argue for melting arc crust and mantle. The high incompatible elements contents with Nd isotope [$\epsilon_{\text{Nd(T)}} = +1.7$ to -2.6] signatures of sanukitoids suggest melting of enriched sub-arc mantle. The high SiO_2 , Nd isotope [$\epsilon_{\text{Nd(T)}} = 0.1$ to -7.3] of anatectic granites suggest reworking of thickened crust through collision of crustal blocks ca. 2560–2500 Ma.

The geodynamic evolution of the Archean crust across the boundary shear zone involves a shift from plume to shallow subduction ca. 3400–3200 Ma and accretionary to collisional orogenesis during ca. 3000–2500 Ma. The rise of continent above sea level by 3000 Ma subjected to oxidative weathering supplied nutrients to oceans. Isotope biomarker data reveal transient surface environments involving anoxic to fluctuating redox conditions and microbial activity by Neoproterozoic. The microbial activity further enhanced biomass production and organic carbon burial, ultimately driving the surface oxygenation led to the building of habitable continent at the dawn of Great Oxygenation Event.

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Contents

1.	Introduction	2
2.	Regional geological and tectonic framework	2
3.	Geological setting	5
3.1.	Western corridor	5
3.1.1.	Granitoid basement (TTG-type gneisses)	5
3.1.2.	Sargur Group greenstone sequences	5
3.1.3.	Bababudan Group - Kibbanahalli arm, Yediyur band and SW of Chitradurga	7
3.1.4.	The Chitradurga Group - Chitradurga greenstone belt	7
3.1.5.	ca. 3000 Ma and 2600 Ma Potassic granite plutons	9
3.1.6.	ca. 3000 Ma Bukkapatna granite	9
3.2.	Chitradurga boundary shear zone	10
3.3.	Eastern corridor	11
3.3.1.	Eastern basement: TTGs and transitional TTGs	11
3.3.2.	Javagondanahalli greenstone belt	12
3.3.3.	Kunigal greenstone belt	13
3.3.4.	Late (2550-2530 Ma) granitoid intrusions in the Eastern basement	13
3.4.	Structural patterns	14
4.	Petrographic characteristics	14
5.	Metamorphic record	15
5.1.	Conventional thermobarometers	17
5.2.	P-T pseudosection analysis	17
5.2.1.	Sample AL-01B - melting of gt-amphibolite	17
5.2.2.	Sample SR19 - retrograded eclogite	18
5.2.3.	Sample 19 CH01	18
6.	Geochronology	19
6.1.	Western corridor	19
6.1.1.	ca. 3400-3180 Ma TTGs	19
6.1.2.	ca. 3000 Ma potassic granites	19
6.1.3.	ca. 2620-2610 Ma potassic granites	19
6.2.	Greenstone volcanics	19
6.2.1.	ca.3400-3150 Ma Sargur Group komatiites	19
6.2.2.	ca. 2900-2850 Ma Bababudan Group mafic volcanics	19
6.2.3.	ca. 2740-2676 Ma Chitradurga Group mafic and felsic volcanics	20
6.2.4.	Bukkapatna granite in the Chitradurga boundary shear zone	20
6.3.	Eastern corridor - (ca. 3300-2500 Ma)	20
6.3.1.	TTGs -Transitional TTGs and granitoid intrusions	20
6.3.2.	Javagondanahalli greenstone belt	20
6.4.	New U-Pb zircon ages of the granitoids: Western corridor	20
6.4.1.	Sample 09 HU-01 - 3313 ± 14 Ma	20
6.4.2.	Sample 18HOL-01B - 3222 ± 7 Ma	20
6.4.3.	Sample 20 WCH-01A - 3237 ± 12 Ma	20
6.4.4.	Sample 20 WCH-06C - 3229 ± 5 Ma	21
6.4.5.	Sample 20 WCH-06A - 3206.1 ± 8.0 Ma	21
6.4.6.	Sample 20 WCH-07A - 3234 ± 8 Ma	21
6.5.	ca. 3000 Ma potassic granite intrusions	21
6.5.1.	Hosdurga pluton: Sample 20 WCH-05B - 2955 ± 5 Ma	21
6.5.2.	Bellur-Nagamangala granite : Sample BC 01 - 2977 ± 7 Ma	21
6.6.	Eastern corridor - TTG-type migmatitic gneisses (3338-3176 Ma)	21
6.6.1.	Sample 19 AL-01A - Granodioritic gneiss: 3338 ± 17 Ma	21
6.6.2.	Sample 20 AL-01A - Tonalitic gneiss: 3258 ± 8 Ma	23
6.6.3.	Sample SB-01 - Granodioritic flat gneiss: 3176 ± 4 Ma	23
6.7.	Transitional TTGs (2669-2618 Ma)	23
6.7.1.	Sample 17CD-A2 - Transitional TTGs: 2633 ± 9 Ma	23
6.7.2.	Sample 18 CD 19 - Transitional TTG: 2638 ± 14 Ma	23
6.7.3.	Sample 17 CD 14 - Transitional TTG: 2640 ± 14 Ma	23
6.7.4.	Sample RAV 01 - Transitional TTG: 2618 ± 9 Ma	23
6.7.5.	Sample 17 CL 01A - Transitional TTG: 2669 ± 7.7 Ma	23
6.8.	Calc-alkaline granitoids (2600-2500 Ma)	26
6.8.1.	Sample 17 JNK 01- Granodiorite : 2597 ± 11 Ma	26
6.8.2.	Sample CD 21- Granite: 2537 ± 8 Ma	26
7.	Chemical characteristics of greenstone – granitoid assemblages	26
7.1.	Greenstone volcanics: major and trace elements	26
7.1.1.	Sargur Group greenstone volcanics (ca.3400-3200 Ma)	26
7.1.2.	Bababudan Group (ca.3000-2720 Ma)	27
7.1.3.	Chitradurga group (ca. 2746-2667 Ma)	28
7.1.4.	Javagondanahalli volcanics (ca. 2740 Ma)	28
8.	Archean granitoids across the boundary shear zone: major and trace element characteristics	29
8.1.	Western corridor - TTG-type granitoids	29
8.1.1.	ca.3000-2950 Ma potassic granites	30
8.1.2.	ca. 2600 Ma high-K granites	31

8.2.	Eastern corridor - ca. 3300-3180 Ma TTG-type gneisses	31
8.2.1.	ca. 2700-2600 Ma transitional TTGs	31
8.2.2.	ca. 2560-2530 Ma sanukitoids	32
8.2.3.	ca. 2520-2513 Ma anatectic granites	32
8.2.4.	ca. 2500 Ma Maddakkanahalli granite	32
9.	Whole rock Nd isotopes	34
9.1.	Greenstone volcanics	34
9.2.	Granitoids (TTGs, transitional TTGs, potassic granites, sanukitoids and anatectic granites)	35
10.	LA-MC-ICM-PS Lu-Hf isotope measurements in zircons	35
10.1.	Western corridor	35
10.2.	Eastern corridor	35
11.	Petrogenesis of greenstone volcanic and granitoids	36
11.1.	Effects of secondary processes on the elemental and isotope compositions	36
11.2.	Origin of greenstone volcanics – <i>Western corridor</i>	36
11.2.1.	Sargur Group komatiite and komatiite basalts	36
11.2.2.	Bababudan Group mafic volcanics (Kibbanahalli arm and Yediyur belt)	36
11.2.3.	Chitradurga volcanics	37
11.3.	Eastern corridor	38
11.3.1.	Javagondanahalli-Kunigal volcanics	38
11.4.	Origin of granitoids: <i>Western corridor</i>	39
11.4.1.	ca. 3350-3200 Ma TTG-type gneisses	39
11.4.2.	ca.3000-2950 Ma potassic granites	40
11.4.3.	ca.2600 Ma potassic granites	43
11.5.	Origin of granitoids: <i>Eastern corridor</i>	44
11.5.1.	ca. 3338-3176 Ma migmatitic TTGs	44
11.5.2.	ca. 2670-2600 Ma transitional TTGs	44
11.5.3.	ca. 2560-2530 Ma sanukitoids	44
11.5.4.	ca. 2507 Ma Maddakkanahalli granite	45
12.	Discussion	45
12.1.	Distinct crustal architecture and magmatic records across the Chitradurga boundary shear zone	45
12.2.	Contrasting metamorphic records in crustal domains across boundary shear zone	45
12.3.	Successive stages of continental growth and mantle evolution: Insights from granitoids and greenstone volcanics	46
12.3.1.	Constraints from zircon U-Pb ages, whole-rock Nd and in-situ Hf isotopes	46
12.3.2.	Constraints from detrital zircon U-Pb ages	47
12.3.3.	Multistage crustal reworking, metamorphism and craton building	48
12.3.4.	Archean surface environments, oxygenation of ocean and emergence of biosphere	49
12.4.	Geodynamic evolution through time and formation of a habitable continent	49
12.4.1.	Multistage accretional and collisional orogenesis	49
13.	Conclusions	50
	CRedit authorship contribution	51
	Declaration of competing interest	51
	Acknowledgements	51
	Data availability	51
	References	51

1. Introduction

Archean cratons form wide time and tectonic windows into early Earth evolution, development of habitable continents, oxygenation of surface envelopes (ocean and atmosphere), and the emergence of a biosphere. Accretionary histories of Archean juvenile crust, craton building processes and transient geodynamics are topics of much discussions in solid-Earth Sciences (Brown et al., 2020; Bedard, 2018; Chardon et al., 2009, 2011; Gerya, 2014; Johnson et al., 2017; Korenaga, 2013; Moyaen and Laurent, 2018; Turner et al., 2013; Polat et al., 2015; Kusky et al., 2018, 2021; Jayananda et al., 2023). Archean cratonic assemblages including TTG suites, volcano-sedimentary greenstone sequences, high potassic granites and sanukitoids provide key chemical probes into crust-mantle evolution, geodynamics of the Archean Earth, origin of cratons and metallogeny (Polat, 2012; Laurent et al., 2014; Smithies et al., 2019; Kamber, 2015; Jayananda et al., 2018, 2023). Numerous petrological, geophysical, geochronologic, elemental and isotope studies together with numerical modelling have provided important constraints on thermo-mechanical processes, petrogenetic mechanisms and tectonics involved in the differentiation of Archean crusts from the mantle (Belousova et al., 2010; Moyaen, 2011; Guitreau et al., 2012; Polat, 2012; Sizova et al., 2015; Baes

et al., 2016; Smithies et al., 2019; Jayananda et al., 2013a; Jayananda et al., 2018). However, geodynamic context of crust accretion and craton building mechanisms are still a subject of much debate more specifically timing of the onset of plate tectonics (Choukroune et al., 1995; Bouhallier et al., 1995; Chardon et al., 1996, 2008, 2011; Brown et al., 2020; Bedard, 2018; Wyman, 2018; Kusky et al., 2021; Ninga et al., 2022; Jayananda et al., 2023). In this context, the Archean Dharwar craton forms an ideal target to address evolving Archean geodynamics, juvenile crust accretion, thermal and chemical evolution of the mantle, reworking and craton building as it displays a large tilted crustal panel forming a wide time (3600-2500 Ma) window (Guitreau et al., 2017; Jayananda et al., 2018, 2020, 2023; Ranjan et al., 2020, 2022). Numerous studies have addressed the magmatism and metamorphic record, tectonic context of crust formation and craton building processes in the Dharwar craton (e.g. Bouhallier et al., 1995; Chadwick et al., 2000; Chardon et al., 1996, 2002, Chardon et al., 2008, Chardon et al., 2011; Gupta et al., 2003; Jayananda et al., 2020, 2023; Krapez et al., 2020; Manikyamba et al., 2017; Vijaya Rao et al., 2015; Mandal et al., 2018; Bidyananda et al., 2016; Dey et al., 2016; Gao and Santosh, 2020; Mohan et al., 2020, 2023; Santosh and Li, 2018; Wang et al., 2023; Ravindran et al., 2023). Recent studies demonstrate that the Archean Dharwar craton is a composite continent comprising collage

of discrete crustal blocks with independent thermal records and crustal accretion histories (e.g. [Peucat et al., 2013](#); [Jayananda et al., 2018, 2020, 2023](#)).

The Chitradurga boundary shear zone has variously been considered as a suture ([Naqvi, 1973](#)), a fundamental terrain boundary separating two crustal blocks ([Drury, 1983](#)), or a transcurrent shear zone ([Chadwick et al., 2000](#)). Moreover, the Chitradurga boundary shear zone separates two major crustal blocks with distinct lithological assemblages, crustal architecture, thermal records, accretionary and reworking histories ([Chardon et al., 2011](#); [Jayananda et al., 2013a, 2013b, 2020](#)). Most of the contributions on these crustal blocks across the boundary shear zone focused either on magmatism (e.g., [Chadwick et al., 2000](#); [Jayananda et al., 2006, 2018, 2020](#); [Mohan et al., 2014, 2023](#); [Ravindran et al., 2023](#); [Sebastian et al., 2021](#); [Mohan et al., 2014, 2023](#)) or strain regimes (e.g., [Chardon and Jayananda, 2008](#); [Chardon et al., 2008](#); [Chardon et al., 2011](#)). However, an integrated study addressing crustal architecture, thermal records, secular changes in petrogenetic processes, geodynamic evolution of crustal blocks across the boundary shear zone as well as redox conditions of surface environments is lacking. In this context, the large crustal corridor across the Chitradurga boundary shear zone forms an extremely relevant target locality to address crustal architecture, thermo-mechanical processes and petrogenetic mechanisms of crustal growth, coupled evolution of crust-mantle system, evolving geodynamics of the Archean Earth and origin of habitable continents. The preserved volcanic rocks of the Sargur Group and Dharwar Supergroup greenstone belts form chemical probes into Archean mantle whilst TTGs, transitional TTGs, sanukitoids and potassic plutons provide constraints on continental growth, secular changes in petrogenesis and geodynamic processes, reworking and stabilization of the Archean crust. Consequently, the main goals of this contribution are to present a comprehensive review of published records in combination with our new field, petrologic data, U-Pb zircon ages combined with in-situ Hf isotopes, whole-rock Nd isotopes, major and trace elements for greenstone volcanics, TTGs, transitional TTGs, granitoid plutons across the Chitradurga boundary shear zone to address: (a) crustal architecture across the boundary shear zone (b) thermal records (c) petrogenesis of the greenstone volcanics, TTG-type basement, potassic plutons and sanukitoids (d) tectonics of micro-blocks including their assembly into stable cratonic framework (e) evolving redox conditions, oxygenation of ocean and microbial activity (f) geodynamics of Archean Earth and origin of habitable continents.

2. Regional geological and tectonic framework

The Archean basement in the Dharwar craton ([Fig. 1a](#)) displays a large tilted section of continental crust displaying a progressive transition from upper to lower crustal levels ([Chardon et al., 2008](#); [Jayananda et al., 2018](#)). The craton contains three major lithological associations including volcanic-sedimentary greenstone assemblages, diverse granitoids including TTG-to transitional TTGs, late calc-alkaline intrusions of sanukitoid affinity and high-potassic granitic plutons. The craton was historically divided into a western and an eastern block which are separated by a north-south trending mylonitic shear zone along the eastern margin of the Chitradurga greenstone belt ([Swami Nath et al., 1976](#); [Chadwick et al., 2000](#); [Chardon et al., 2008, 2011](#)). Subsequent studies identified three crustal blocks (western, central and eastern) with distinct crustal architectures, rock assemblages, thermal records and accretionary histories ([Jayananda et al., 2013a](#); [Peucat et al., 2013](#)). The western block preserves an ancient granitoid basement (ca. 3600–3170 Ma) with two generations of greenstone sequences and ca. 3000 Ma and 2600 Ma high-potassic granite plutons ([Guitreau et al., 2017](#); [Jayananda et al., 2015, 2020, 2023](#)). The greenstone assemblages comprise older (>3200 Ma) Sargur Group and younger (<3000 Ma) Dharwar Supergroup both of which are separated by angular unconformity documented at the base of Bababudan, Sigegudda and Kibbanahalli arm ([Ramakrishnan et al., 1976](#); [Chadwick et al., 1981](#); [Venkata](#)

[Dasu et al., 1991](#); [Chardon et al., 1996](#); [Krapez et al., 2020](#)). The older Sargur Group comprises ca. 3400–3200 Ma komatiite-dominated volcanics with subordinate high-Mg basalt to tholeiitic basalts and minor pyroclastic felsic flows with interlayered sedimentary rocks (quartzite-pelite-carbonate-BIFs). Detrital U-Pb zircon ages indicate ca. 3600–3200 Ma ([Nutman et al., 1992](#); [Hokada et al., 2013](#); [Bidyananda et al., 2016](#); [Sarma et al., 2012](#); [Lancaster et al., 2015](#); [Wang and Santosh, 2019](#); [Ranjan et al., 2022](#); [Mittra et al., 2023](#); [Wang et al., 2023](#)). The ca. 3400–3200 Ma Sargur Group greenstone assemblages and adjacent TTGs are intruded by ca. 3000–2950 Ma potassic plutons ([Jayananda et al., 2019](#); [Ao et al., 2021](#); [Sebastian et al., 2021](#); [Corfu and Hegde, 2020](#)). The younger Dharwar Supergroup is further sub-divided into a lower Bababudan Group and an upper Chitradurga Group ([Chadwick et al., 1981](#); [Swami Nath and Ramakrishnan, 1981](#)). The Bababudan Group in the type locality begins with oligomictic quartz-pebble conglomerate/quartzite followed by at least three thick basaltic flows with quartzite layers, phyllite, felsic volcanics and finally thick BIFs at the summits ([Chadwick et al., 1985](#)). Whole-rock Sm-Nd whole rock isochron ages of mafic volcanics together with U-Pb zircon ages of tuffs indicate ca. 3000–2720 Ma ([Drury, 1983](#); [Kumar et al., 1996](#); [Trendall et al., 1997](#)) for the timing of volcanism. Detailed strain fabrics mapping, and kinematic analysis of the Bababudan Group reveal centripetal collapse of high-density greenstone load into low density basement ([Chardon et al., 1996, 1998](#)). The Chitradurga Group comprises tholeiitic basalts, intermediate to felsic volcanic rocks associated with shallow to deep water sedimentary assemblages which in turn intruded by the potassic plutons ([Seshadri et al., 1981](#)). Geochronological data indicate ca. 2740 – 2676 Ma age ([Kumar et al., 1996](#); [Nutman et al., 1996](#); [Jayananda et al., 2013b](#); [Hokada et al., 2013](#)) for basaltic to pyroclastic felsic flows whilst U-Pb detrital zircon ages indicate ca. 3400 – 2600 Ma ([Hokada et al., 2013](#); [Wang et al., 2019](#); [Sarma et al., 2012](#); [Nasheeth et al., 2016](#); [Krapez et al., 2020](#)). Late potassic plutons emplaced into the ca. Chitradurga- Shimoga basins and adjoining basement during a narrow time window of 2615–2590 Ma ([Jayananda et al., 2006](#); [Mohan et al., 2014, 2023](#); [Tarun Kumar et al., 2022](#)). The western block experienced an early thermal event at ca. 3100 Ma which was immediately preceded by emplacement of ca. 3200–3150 Ma diapiroic trondhjemite plutons ([Jayananda et al., 2013a, 2015, 2023](#); [Dasgupta et al., 2019](#)), and a terminal ca. 2500 Ma regional thermal event that affected the entire Archean crust in the Dharwar craton.

The central block exposes a granitoid basement composed of voluminous younger (ca. 2700–2600 Ma) banded transitional TTGs with large discontinuous outcrops of older (ca. 3360–3000 Ma) TTG-type migmatitic grey gneisses ([Peucat et al., 2013](#); [Bidyananda et al., 2016](#); [Chardon et al., 2011](#); [Jayananda et al., 2000, 2020](#); [Mohan et al., 2023](#); [Ratheesh Kumar et al., 2016](#)). The greenstones sequences are dominated by tholeiitic to alkaline basalts with minor felsic to intermediate volcanics ([Manikyamba et al., 2012, 2017](#)) which show bimodal age distribution including ca. 2700 Ma ([Balakrishnan et al., 1999](#); [Naqvi et al., 2002](#); [Jayananda et al., 2013b](#); [Khanna et al., 2016](#)) and 2570–2545 Ma ([Nutman et al., 1996](#); [Jayananda et al., 2013b](#)). North – South trending calc-alkaline to potassic plutons are composite and comprise voluminous juvenile granitoids of a sanukitoid affinity and minor anatectic granites derived from pre-existing basement ([Jayananda et al., 1995, 2000, 2020](#); [Gireesh et al., 2012](#); [Dey et al., 2016](#); [Mohan et al., 2013](#)). The central block experienced three thermal events ca. 3200 Ma amphibolite facies and 2620 Ma UHT event ([Jayananda et al., 2011, 2013a](#); [Kaempfer et al., 2024](#)) and 2500 Ma ([Jayananda et al., 2013a](#); [Peucat et al., 2013](#)).

The eastern block contains younger (ca. 2700–2550 Ma) granitoid basement rocks ([Krogstad et al., 1991](#); [Peucat et al., 1993, 2013](#); [Jayananda et al., 2020](#); [Dey et al., 2016](#); [Wang et al., 2020](#)) of a transitional-TTG affinity, thin elongated volcanic-dominated (minor komatiite-tholeiite-boninite- felsic pyroclastic flows) greenstone belts with bimodal age distribution of ca. 2700 Ma and ca. 2550 Ma ([Dey et al., 2015, 2018](#); [Jayananda et al., 2013b](#); [Khanna et al., 2016](#); [Khanna,](#)

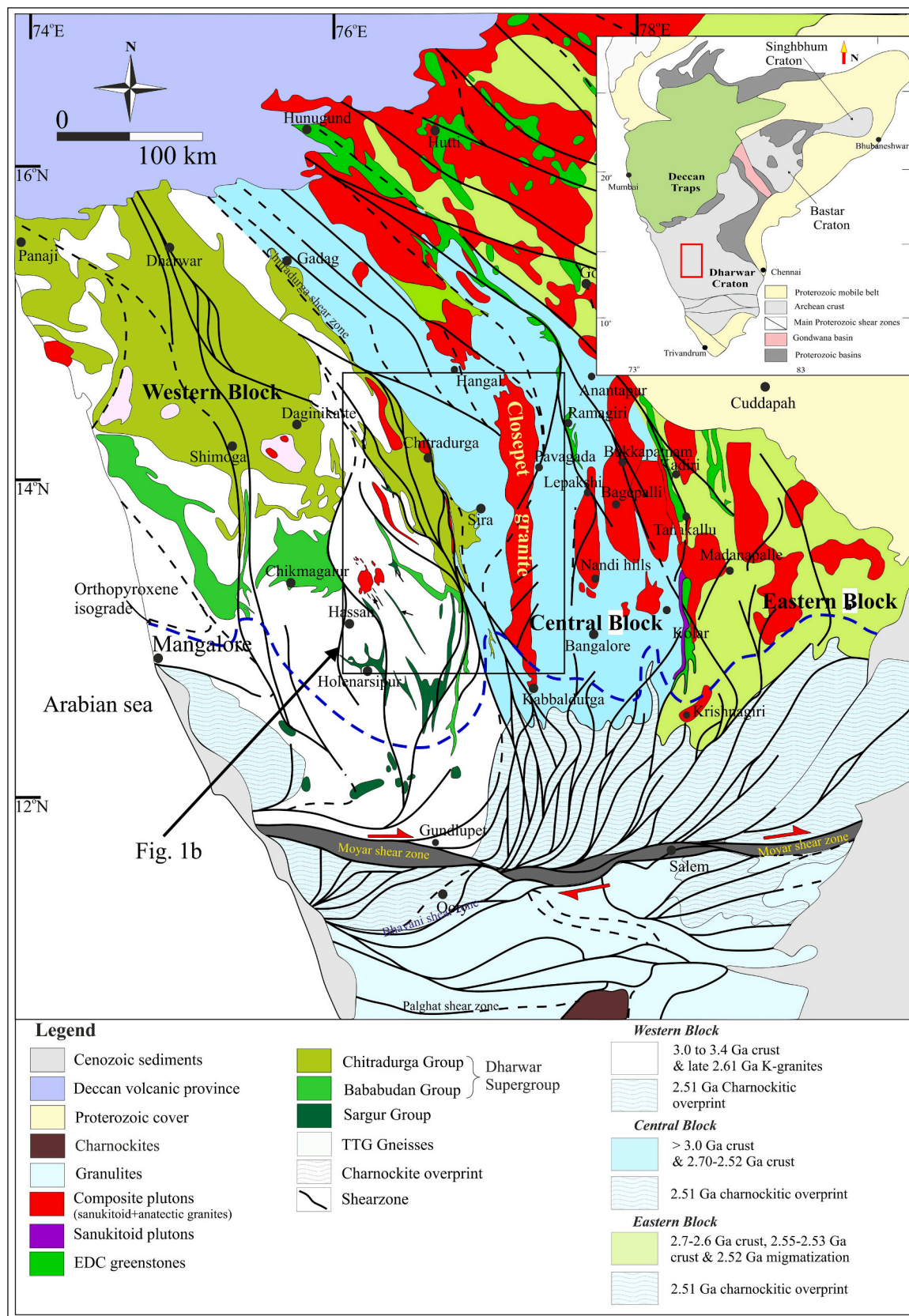
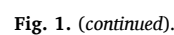


Fig. 1. (a) Geological sketch map of the Dharwar craton (modified after [Peucat et al., 2013](#)) showing rectangular box depicting the present study area. Inset map of peninsular India showing the Dharwar craton in the rectangular box (b) Geological sketch map of the studied corridor (modified after [Jayananda et al., 2018](#)).



thermal event close to 2500 Ma that affected the entire Archean crust in the Dharwar craton (Peucat et al., 2013; Jayananda et al., 2013a). The three crustal blocks assembled into a cratonic framework at 2560-2500

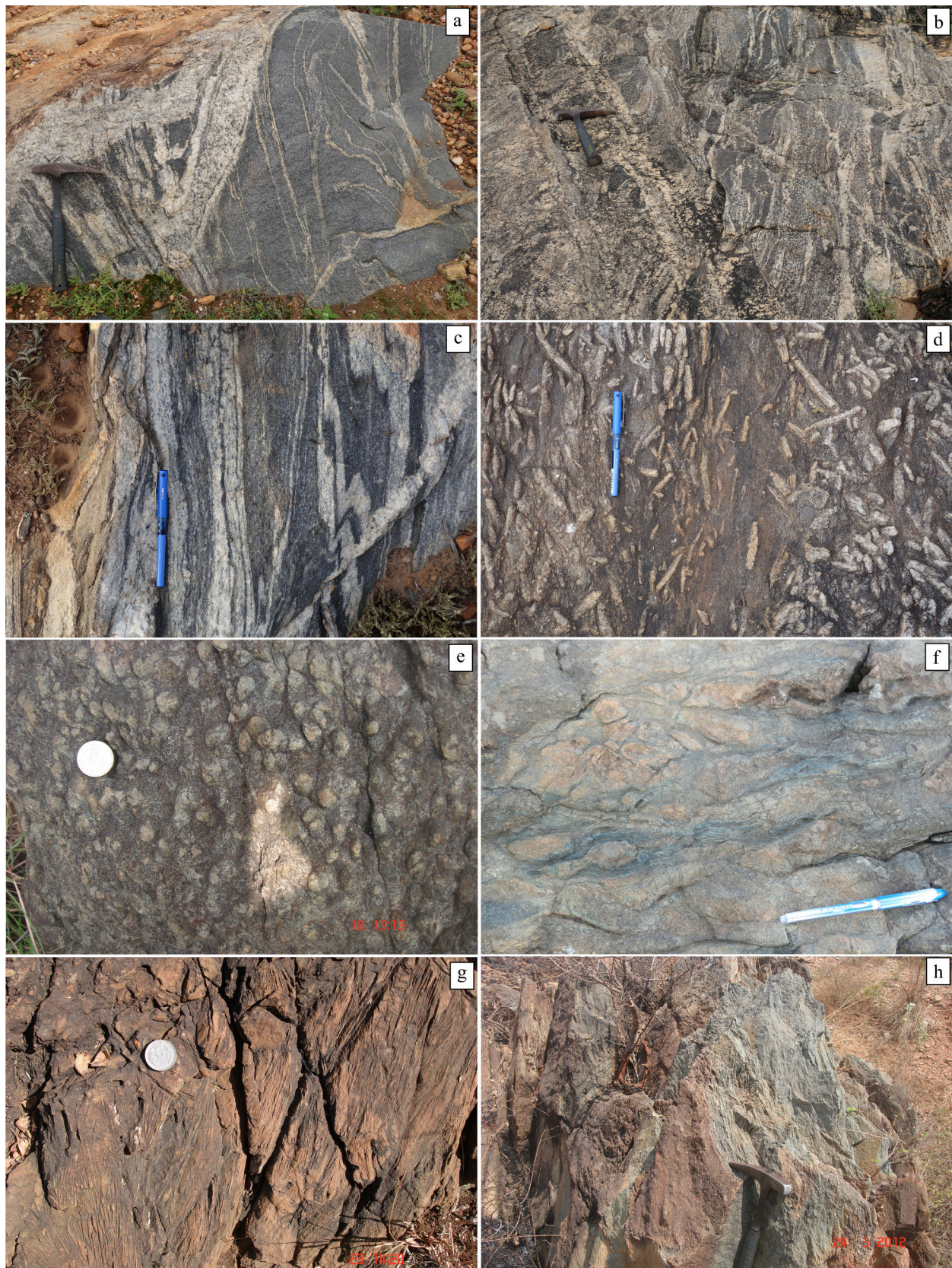


Fig. 2. (a) TTG-type gneisses show dark grey tonalitic to grey granodioritic facies both of which traversed by whitish grey trondhjemite (b) Weakly foliated TTG gneiss traversed by highly coarse-grained pegmatite dyke (c) TTG gneiss from the western margin of the Chitradurga belt (d) Discontinuous outcrops of pelitic rock with large kyanite-crystals from the western margin of Chitradurga belt (e) Nodular structures in komatiites from the J.C. Pura greenstone belt (f) Flow top breccia from the Banasandra greenstone belt (g) Komatiite showing spinifex structure from the Banasandra greenstone belt (h) Deformed pillows in the komatiite from the Nagamangala greenstone belt.

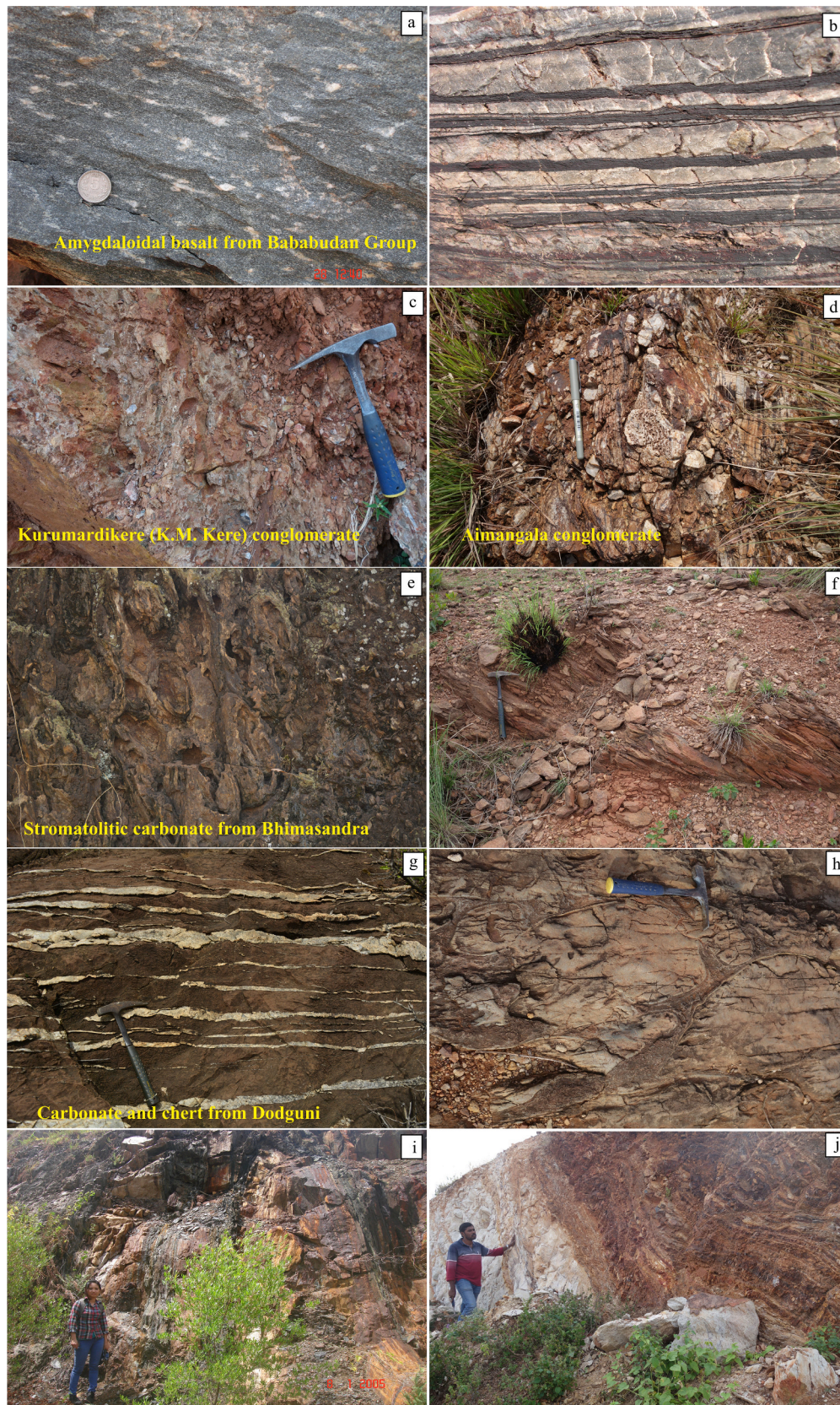


Fig. 3. (a) Amygdaloidal basalt from Bababudan Group Kibbanahalli arm (b) Spectacular BIF with alternate layers of chert and iron oxide from Kibbanahalli arm (c) Kurumardikere (K.M. Kere) conglomerate in the central part of the Chitradurga greenstone belt (d) Aimangala conglomerate with diverse angular clasts at the contact between Chitradurga belt and Javagondanahalli belt (e) stromatolitic carbonate with concentric rings of preserved algal mats along the western margin of Chitradurga belt near Bhimasamudra village (f) siliciclastics along the south western margin of the studied corridor in the Chitradurga belt (g) carbonates with chert

layers in the Dodguni area (h) Basalt displaying large pillows with chilled margins at about 5 km west of Burujana Roppa (i) carbonaceous shale interbedded with sulphide bearing BIFs at Ajjanahalli mines (j) greywacke-argillite-ash beds interlayered.

Ma through westward convergence of orogenic crust (Chadwick et al., 2000; Chardon et al., 2011; Jayananda et al., 2013b, 2020; Vijaya Rao et al., 2015).

3. Geological setting

The Chitradurga greenstone belt (Seshadri et al., 1981) runs at least 450 km in broadly north-south direction from Srirangapatna to Gadag with NNE strike in the South whilst NNW strike to the North (Fig. 1b). A major shear zone traverses the eastern margin of the Chitradurga greenstone belt that separates two crustal blocks corresponding to different age provinces with distinct crustal histories. In the south-central to central part, the Chitradurga boundary shear zone (at the latitude of Sira-Chitradurga, see Fig. 1b) separates two volcano-sedimentary units which comprise Chitradurga greenstone belt to the west and Javagondanahalli greenstone belt to the east (see Fig. 1b) where the shear zone contains ca. 3000 Ma old mylonitic Bukkapatna granite (Chardon et al., 2011) and adjoining greenstone units whilst to the north of Hiriyur, outcrops of Bukkapatna granite are rarely exposed on the surface, but thin band of north-south trending polymictic Aimangala conglomerate separate Chitradurga and Javagondanahalli greenstone belts (see Fig. 1b). Further, to the north of Aimangala conglomerate, along the eastern margin of the Chitradurga belt deformed granites yielded U-Pb zircon ages of 2992 ± 6 Ma and 2962 ± 11 Ma (Mohan et al., 2023) which coincides with U-Pb zircon age of the Bukkapatna granite (Chardon et al., 2011). About 80 km south of Sira, the Kunigal greenstone belt (Fig. 1b) shows lithological assemblages similar to the Javagondanahalli belt. In the following sections, we outline the geological framework of crustal domains across the Chitradurga boundary shear zone covering a large E-W corridor (about 100 km width) from Chitradurga town in the north and Kunigal in the south (Fig. 1b).

3.1. Western corridor

The crustal corridor to the west of the boundary shear zone comprises Archean basement (TTG-type granitoids with Sargur Group), younger cover sequences of the Bababudan and Chitradurga Group volcanic-sedimentary greenstone assemblages. Late high potassic granite intrusions include elongated ca. 3000 Ma Bellur-Nagamangala granite and Hosdurga plutons, and ca. 2600 Ma Chitradurga, Arsikere-Banavara and Pandavapura plutons (see Fig. 1a).

3.1.1. Granitoid basement (TTG-type gneisses)

The granitoid basement along the western part of the Chitradurga greenstone belt contains TTG-type gneisses composed of dark grey tonalitic to grey granodioritic facies both of which traversed by whitish grey trondhjemite (Fig. 2a). The tonalitic to granodiorite facies appears to be coeval as they display diffuse contacts whilst trondhjemite is a later injection found along the foliation. All the three facies are traversed by highly coarse-grained pegmatite dykes. Disrupted mafic layers or lenses can be found within TTGs along the foliation (Fig. 2b). The TTGs are banded to weakly migmatitic and show shallow to steep foliation (Fig. 2c). High strain zones marked by N30-32°W sinistral shears traverse the gneisses along the western boundary of Chitradurga greenstone belt. Further southwest in the Tiptur - J.C. Pura region these TTGs form sub-circular domes whilst adjoining komatiite-komatiitic basalts coincide with keels (Chardon et al., 1996).

3.1.2. Sargur Group greenstone sequences

A series of well-preserved komatiite-dominated volcano-sedimentary sequences belong to the Sargur Group (Jayananda et al., 2008) located

along the western boundary of the Chitradurga greenstone belt. These greenstone sequences include the Ghattihosahalli, J.C. Pura, Banasandra and Nagamangala greenstone belts apart from numerous disrupted fragments found within the TTG-type granitoid basement (see Fig. 1b).

Viswanatha et al. (1977) first reported spinifex textured komatiites from Ghattihosahalli belt and subsequently Jayananda et al. (2008) documented petrological characteristics. The Ghattihosahalli greenstone belt (see Fig. 1b) is an NNW trending band which comprises disrupted komatiite dominated volcanics interlayered with fuchsite quartzite with barite beds. Komatiites can be found as bouldery strewn outcrops with remnants of spinifex texture. Further north, around Kudineerkatte village pelitic rocks with large kyanite-crystals (Fig. 2d) are found interlayered with amphibolites. About 8 km NNW of Huliyaar in Banavara road, several NNW trending discontinuous outcrops of fine grained greenish grey komatiites to komatiitic basalts. These show evidence of shear deformation as evidenced by massive to schistose fabrics with steep foliation.

The J.C. Pura greenstone belt (see Fig. 1b) is composed of well-preserved abundant ultramafic rocks of komatiite lineage and minor mafic rocks found in association with thin cherty quartzite and BIFs layers. The details of geology and stratigraphic relationships reveal that J.C. Pura greenstone sequences are overlain by younger Kibbanahalli arm (Bababudan Group) with a stratigraphic break (Venkata Dasu et al., 1991). The ultramafic rocks are primarily of massive and schistose peridotitic komatiites, serpentinites, locally talc-tremolite-chlorite-actinolite schists, magnetite-chlorite-talc schists whilst mafic rocks are actinolite-chlorite-plagioclase bearing. The komatiites and komatiite basalts are massive greenish grey to greyish coloured with occasional spinifex textures but generally form spectacular pillows to flow top breccias and ocelli structures which erupted in multiple pulses (Jayananda et al., 2016). Nodular or ocelli structures consist of spherical nodules or ocelli 1-2 cm diameter (Fig. 2e) in fine grained serpentine matrix, which was interpreted as products of liquid immiscibility or chilling process during extrusion (Jayananda et al., 2016). Komatiites and komatiitic basalts show varying degree of serpentinization and carbonation. Sm-Nd whole-rock isochron define an age of 3384 ± 200 Ma for these komatiites (Jayananda et al., 2008).

The Banasandra greenstone belt (see Fig. 1b) occurs as NNW trending and elongated north-south trending discontinuous bands (Srikantia and Bose, 1985; Jayananda et al., 2008, 2016). Details of physical volcanology and petrological characteristics of the komatiite flows were documented by Jayananda et al. (2016). Komatiites are the most voluminous rock type and they occur as fine-grained massive to bouldery outcrops with cumulate facies comprising serpentinite, dunite and flows in the higher levels showing pillow structures, pillow breccia to flow top breccia or hyaloclastites in the southern part of the belt (Fig. 2f). They show spectacular spinifex structure with olivine needles measuring up to 12-13 mm at the lower levels and progressively smaller needles at the higher levels documented in the northern part (Fig. 2g). Sm-Nd whole-rock isochron defined by samples of Banasandra, Ghattihosahalli and J. C. Pura volcanics indicate a 3356 ± 110 Ma age (Jayananda et al., 2008) whilst Maya et al. (2017) obtained a Sm-Nd whole rock isochron age of 3136 ± 200 Ma for komatiite samples of the Banasandra belt.

The Nuggihalli greenstone belt is composed of abundant ultramafic rocks (peridotite to serpentinites), actinolite-tremolite-talc schists with minor amphibolites and interlayered metasediments. The ultramafic rocks rarely preserve primary mineralogy of olivine and pyroxene. Jaffri et al. (1997) have described spinifex textured komatiite in the northern part of the belt. The greenstone sequence contains chromitite layers spatially associated with the serpentine rich ultramafic rocks (Ma et al., 2024). Komatiites from Nuggihalli and stratigraphically equivalent adjacent Kalyadi belt yield Sm-Nd whole-rock isochron age of $3284 \pm$

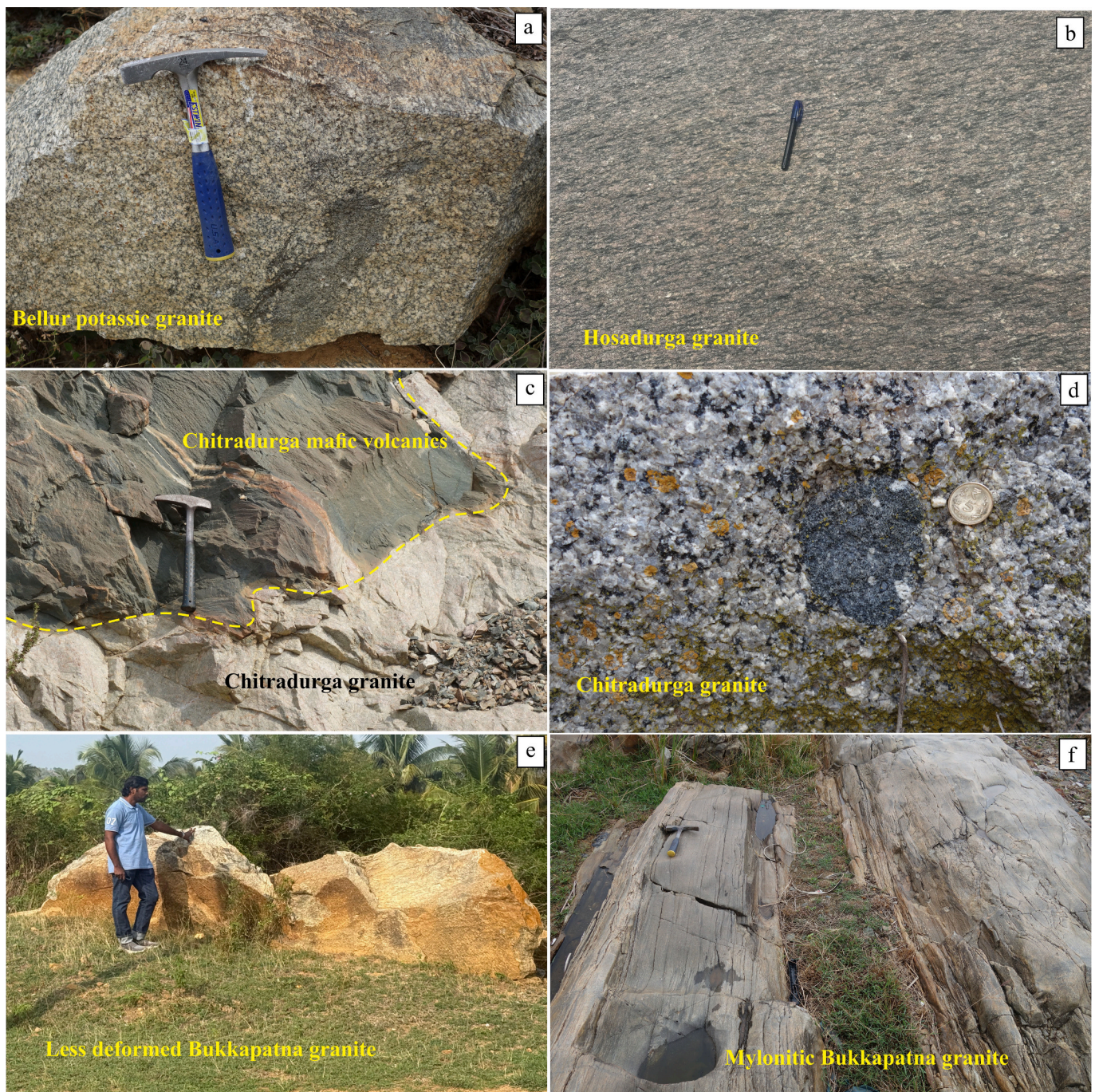


Fig. 4. (a) ca. 3000 Ma Bellur potassic granite contains large pillowed mafic magmatic enclave (b) ca. 3000 Ma old Hosadurga granite showing strong north-south trending fabrics (c) Chitradurga granite intrude the greenstone volcanics at the highway roadcut about 5 km east of Chitradurga town (d) Highly coarse grained Chitradurga granite containing circular mafic magmatic enclaves (e) ca. 3000 Ma old mylonitic Bukkapatna granite showing vertical foliation with horizontal to shallow lineation (f) Mylonitic Bukkapatna granite.

310 Ma (Jayananda et al., 2008) whilst Mukherjee et al. (2012) obtained a whole-rock Sm-Nd isochron age of 3150 ± 120 Ma for komatiites of Nuggihalli belt.

The Nagamangala greenstone belt is located in the SW margin of the Chitradurga greenstone belt (see Fig. 1b) and it extends from Bellur in the North to Honnakere in the South (Srikantia and Venkataramana, 1989). The greenstone belt is surrounded by ca. 3300-3200 Ma TTG basement and intruded by ca. 3000 Ma potassic plutons (Jayananda et al., 2019; Ao et al., 2021). The lithological assemblages include voluminous ultramafic volcanic rocks with minor mafic volcanic rocks and interlayered sedimentary rocks like quartzites, pelite and BIFs. The

whitish green to greenish grey coloured ultramafic rocks comprises massive to schistose serpentinites, talc-tremolite, talc-tremolite-actinolite bearing komatiites (Tushipokla and Jayananda, 2013). In several outcrops, remnants of mesoscopic structures like randomly oriented actinolite crystals have been interpreted as spinifex textures (Devapriyan et al., 1994). However, these pseudo-spinifex structures could have developed during post-magmatic secondary processes (Tushipokla and Jayananda, 2013). Spinifex-textured komatiites with criss-crossing sheafs of closely spaced needles (3-7 cm length) of olivine (mostly converted into serpentine) can be found in the southeastern part of the belt. Strongly deformed pillows can be occasionally observed

north of Honnakere (Fig. 2h) and further north (Srikantia and Venkataramana, 1989; Tushipokla and Jayananda, 2013). The mafic rocks include hornblende schist and amphibolite and are confined to higher stratigraphic levels and show well preserved crude pillow structures with chilled margins. The ultramafic to mafic rocks exhibit steep foliation/cleavage and are traversed by thin quartz veins. Spectacular outcrops of metapelites with large kyanite crystal concentration can be found near Nelligere village.

3.1.3. Bababudan Group - Kibbanahalli arm, Yediyur band and SW of Chitradurga

In the studied crustal corridor, volcano-sedimentary sequences belonging to the Bababudan Group (Swami Nath and Ramakrishnan, 1981) are preserved in the area southwest of the Chitradurga town, the Kibbanahalli arm (Venkata Dasu et al., 1991) and Yediyur belt in the south.

The Kibbanahalli arm (see Fig. 1b) extends about 50 km in NW direction from Kibbanahalli village to Vedavathi River (Swami Nath et al., 1974). The lithological assemblages include oligomictic conglomerates/quartzite, mafic to intermediate volcanic rocks and BIFs. The mafic rocks comprise amygdaloidal basalts (Fig. 3a) that were affected by greenschist to epidote amphibolite facies metamorphism. The BIFs are confined to summit which show spectacular chert and iron oxide layers (Fig. 3b). The Kibbanahalli arm overlies the Sargur Group J.C. Pura greenstone belt with decollement defined by quartzite/conglomerate (Venkata Dasu et al., 1991; Chardon et al., 1996). Detailed structural analysis reveals two strain fields (D1 and D2) associated with two greenstone cycles where D1 strain affects only the older Sargur Group greenstones and adjoining TTGs compatible with gravitational

unstabilities that produced dome-basin pattern. In contrast, D2 deformation affects younger Kibbanahalli arm which defines the overall geometry of a syncline that is discordant with respect to older J.C. Pura greenstone and adjoining TTGs (Chardon et al., 1996). The base of the Kibbanahalli arm is tectonized and corresponds to a decollement surface with overall strain patterns compatible with the development of sagducting trough (Chardon et al., 1996) whilst other workers interpreted it as a failed rift (Sreehari et al., 2021).

The N-S trending thin Yediyur band (2-3 km wide) forms part of the southern part of the Chitradurga greenstone belt (see Fig. 1b.) belongs to the Bababudan Group (Swami Nath and Ramakrishnan, 1981). This thin greenstone unit (corresponds to a high-strain zone where contacts with the adjoining granitoid basement are tectonized. The greenstone sequences comprise fine-grained highly schistose dark grey hornblende-actinolite bearing amphibolites, phyllite-argillite-carbonaceous meta-shale and BIFs. All the rocks show strong fabrics with thin banding and steep cleavage with tight folds in argillite and carbonaceous shale.

In the northern part of the studied corridor (see Fig. 1b), the Bababudan Group volcano-sedimentary greenstone assemblages are confined to the southwest of Chitradurga town (see Fig. 1b) and are preserved a NW-SE trending thin band from Neralakatte to the latitude of Siranakatte dome in the southeast. The rock assemblages include oligomictic quartz-pebble conglomerate, current bedded quartzites, amygdular basalts and BIFs (Seshadri et al., 1981; Chadwick et al., 1981) which in turn are overlain by the younger Chitradurga Group.

3.1.4. The Chitradurga Group - Chitradurga greenstone belt

Detailed field-based studies involving lithological mapping and structural analysis have provided a first order geological, stratigraphic

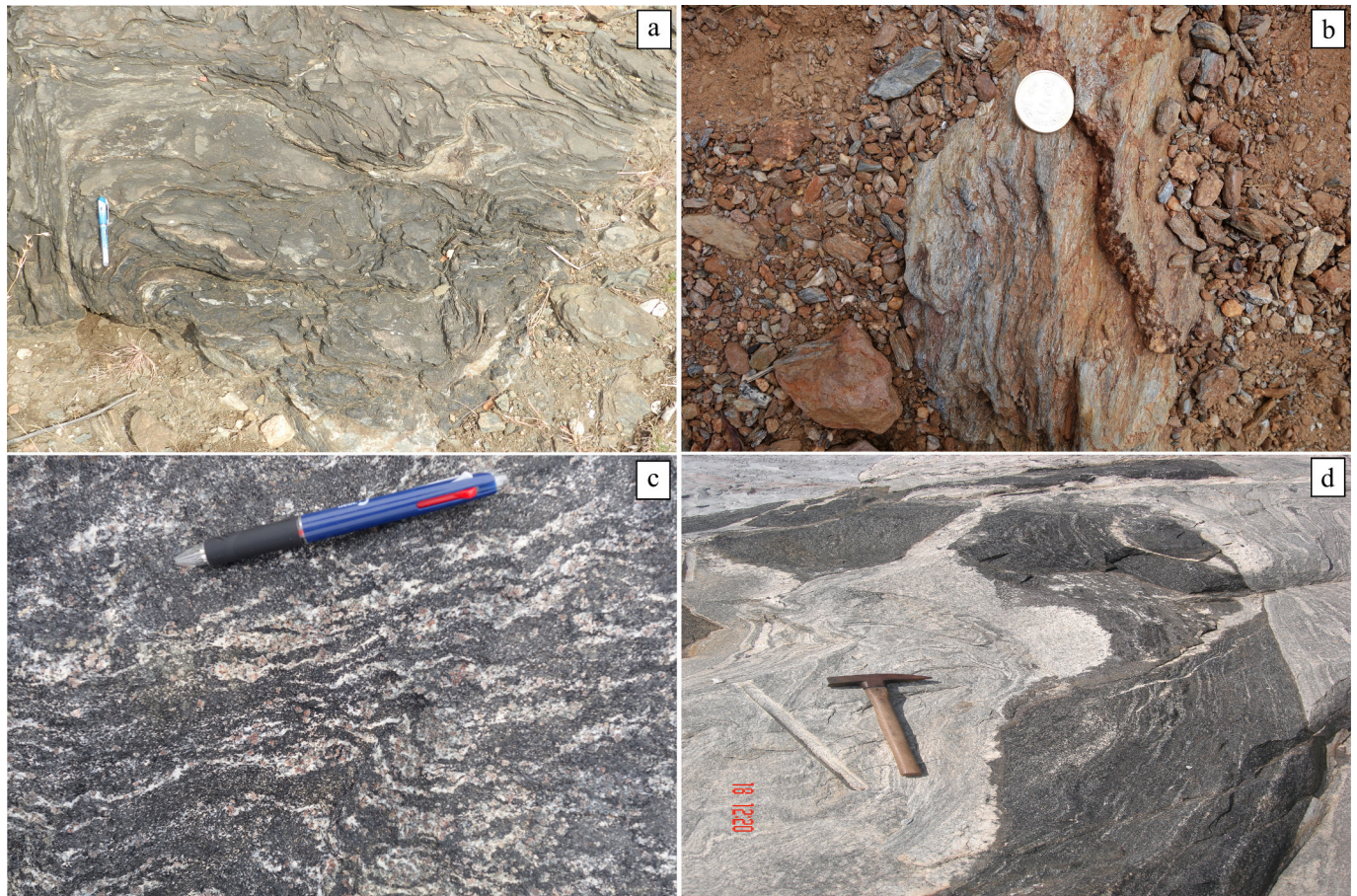


Fig. 5. (a) Javagondanahalli greenstone mafic volcanics showing shallow fabrics (b) mica schist from high-strain zone (c) melting amphibolite with garnet both in leucosomes and mesosomes (d) large fragment of melted garnet-amphibolite within the migmatitic TTG gneiss.

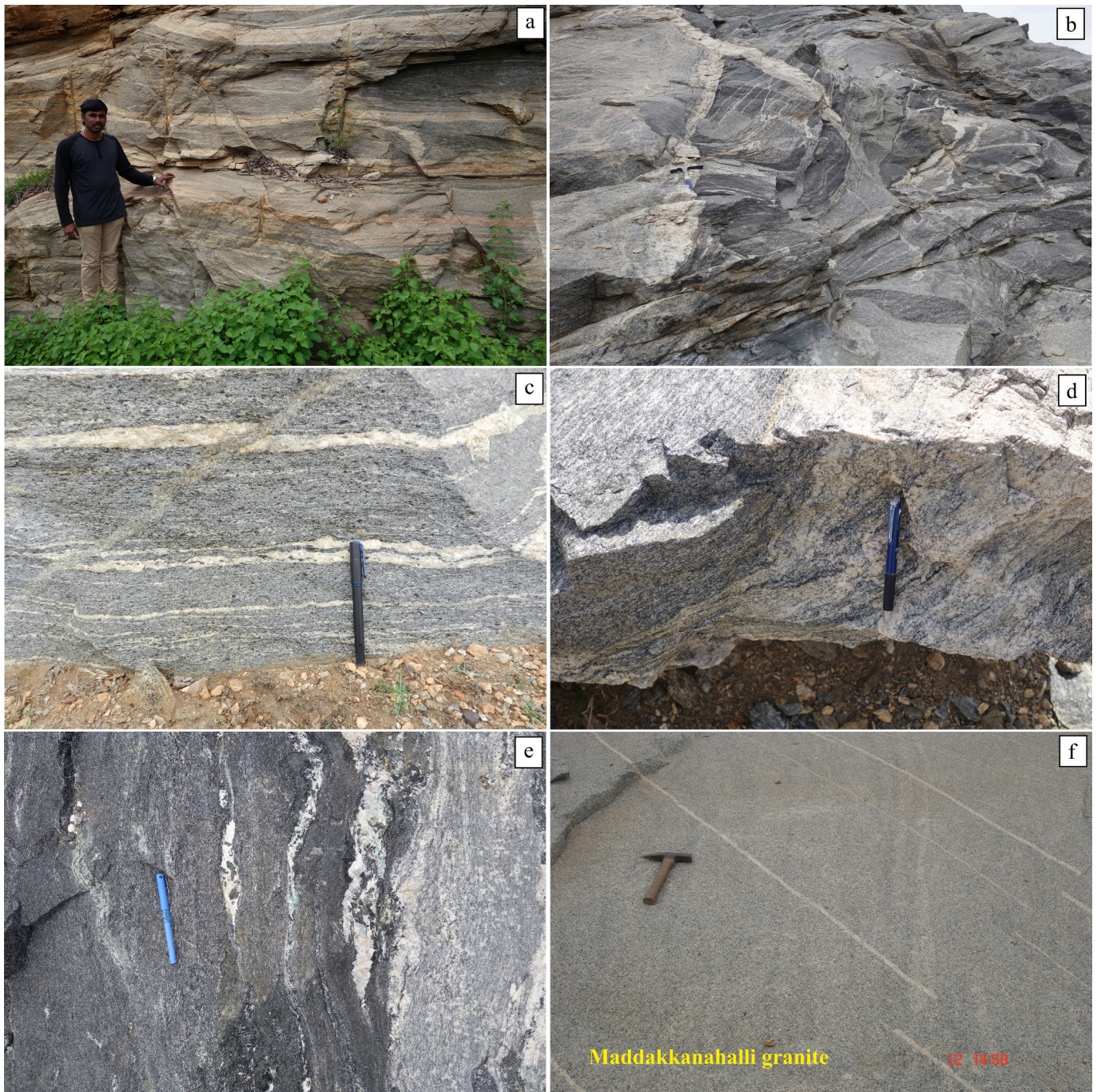


Fig. 6. (a) TTG gneiss show flat fabrics containing pink vein along flat foliation (b) TTG gneiss traversed by ductile shear zones where the flat fabrics are transposed into steeper shears (c) Transitional TTGs show flat to shallow fabrics (d) Transitional TTG show diffused contact with older migmatitic gneisses and both show flat to steep fabrics (e) Transitional TTG and mafic rock show steep fabrics where mafic rock traversed by tiny leucocratic veins containing clinopyroxene and fluorite crystals (f) aplitic Maddakkanahalli granite intrude the contact between Javagondanahalli greenstone belt and Transitional TTGs.

and tectonic framework of Chitradurga greenstone belt (Naqvi, 1973; Harinadha Babu et al., 1981; Seshadri et al., 1981; Chadwick et al., 1981). Subsequently, a tectonic fabric analysis of the Chitradurga greenstone belt and adjoining granitoids have documented two episodes of shear deformation (Jayananda et al., 2006; Chardon et al., 2008). More recent studies provided details about the stratigraphy, structure and tectonic framework of the Chitradurga greenstone belt and adjoining basement rocks (Krapez et al., 2020; Roy et al., 2020).

The Chitradurga greenstone belt is a composite terrain that comprises two distinct volcano-sedimentary sequences corresponding to two tectono-stratigraphic units (i.e. older Bababudan Group in the western

part and younger Chitradurga Group). Both the stratigraphic units are separated by polymictic Talya conglomerate (Chadwick et al., 1981; Ojakangas et al., 2014) in the west which forms the base for the Chitradurga Group. The eastern margin of Chitradurga greenstone belt is bounded by a major crustal scale boundary shear zone containing ca. 3000 Ma mylonitic Bukkapatna granite (Swami Nath et al., 1976; Chadwick et al., 1989; Jayananda et al., 2006; Chardon et al., 2011). The shear zone is bounded by volcanic-dominated Javagondanahalli greenstone belt to the east. We present field relationships of the Chitradurga greenstone belt along three major NE-SW corridors in the north-central part at the latitude of the Chitradurga town to Holalkere

(see Fig. 1b) and a corridor from Ajjanahalli gold mines to Kibbanahalli cross (K.B. Cross). The southern part of the study area corresponds to a E-W corridor from Kunigal to Bellur cross (see Fig. 1b).

The Chitradurga Group volcano-sedimentary sequences in the type locality (SW of Chitradurga) unconformably overlies the granitoid basement and Bababudan Group. The Bababudan Group sequences in turn overlies the basement (granitoids and associated Sargur Group greenstone sequences) and comprises oligomictic Neralakatte conglomerate and basaltic flows with phyllite and BIFs.

The Chitradurga greenstone sequence contains three conglomerate horizons including Talya conglomerate in the west, Kurumardikere (K. M. Kere) conglomerate (Fig. 3c) in the central part (see Fig. 1b), and Aimangala conglomerate (Fig. 3d) to the east. These conglomerates may probably form marker horizons for the three lithostratigraphic and tectonic units which is consistent with the documentation of three distinct BIF units (BIF I, BIF II, BIF III) oldest to youngest from west to east respectively (Chadwick et al., 1981). Our field data including the nature of pebbles, and the matrix coupled with associated rock assemblages and published detrital zircon age data (Krapez et al., 2020; Nasheeth et al., 2016) imply distinct provenance and tectonic environments.

In the northern part of studied corridor, the Chitradurga greenstone succession commences with basal polymictic Talya conglomerate followed by siliceous phyllites, siliciclastic layers, carbonates, metamorphosed mafic volcanics and BIFs. The Talya conglomerate contains pebbles of diverse assemblages which often do not have a clast supported framework as they are embedded in clays. Ojakangas et al. (2014) considered that Talya conglomerate is of glacio-marine origin. Carbonates in the outskirts of Bhimasamudra village (Fig. 3e) are

stromatolitic with well-preserved concentric rings of algal mats. East of Talya conglomerate horizon, basaltic rocks associated with siliciclastic rocks rarely contain garnet (Fig. 3f) whilst BIFs contain alternate layers of chert and iron oxides of variable thicknesses which were affected by superimposed deformation associated with regional shear deformation.

The studied northern corridor is stratigraphically divided into three formations; Lower Vani Vilas Formation, middle Ingldhal Formation and upper Hiriyur Formation (Seshadri et al., 1981) which was subsequently adopted by other workers (Hokada et al., 2013; Ojakangas et al., 2014; Krapez et al., 2020). The Vani Vilas Formation contains quartzite, oligomictic to polymictic conglomerates, carbonates and BIFs I (Seshadri et al., 1981; Harinadha Babu et al., 1981) sequences that are confined to the western part of the basin suggesting stable shelf environments to tectonically unstable conditions. Further southwest, rock assemblages of Vani Vilas Formation include quartzites, stromatolitic cherty dolomite, alternate layers of carbonate and chert layers found in Dodguni- Chikknayakanahalli section (Fig. 3g) reveal shallower continental shelf environment with periodic fluctuations in oxygen on ocean surface wherein carbonate precipitate consuming the available oxygen whilst chert precipitated during oxygen poor conditions due to abundant accumulation of organics (c.f. Hashizume et al., 2016). These carbonate sequences often form typical stromatolites with concentric structures (see Fig. 3e). The Ingldhal Formation are mainly confined to the central part which is dominated by mafic to felsic volcanics with BIFs II (Chadwick et al., 1981) that are associated with basaltic flows around the Bellara – Ajjanahalli region and to the northeast (Chitradurga, Ingldhal, Jogi Mardi, Kallehadlu and Burujana Roppa region) where fine-grained basalts form large circular pillows with chilled margins (Fig. 3h). Intermediate to felsic volcanics occurs as massive pyroclastic

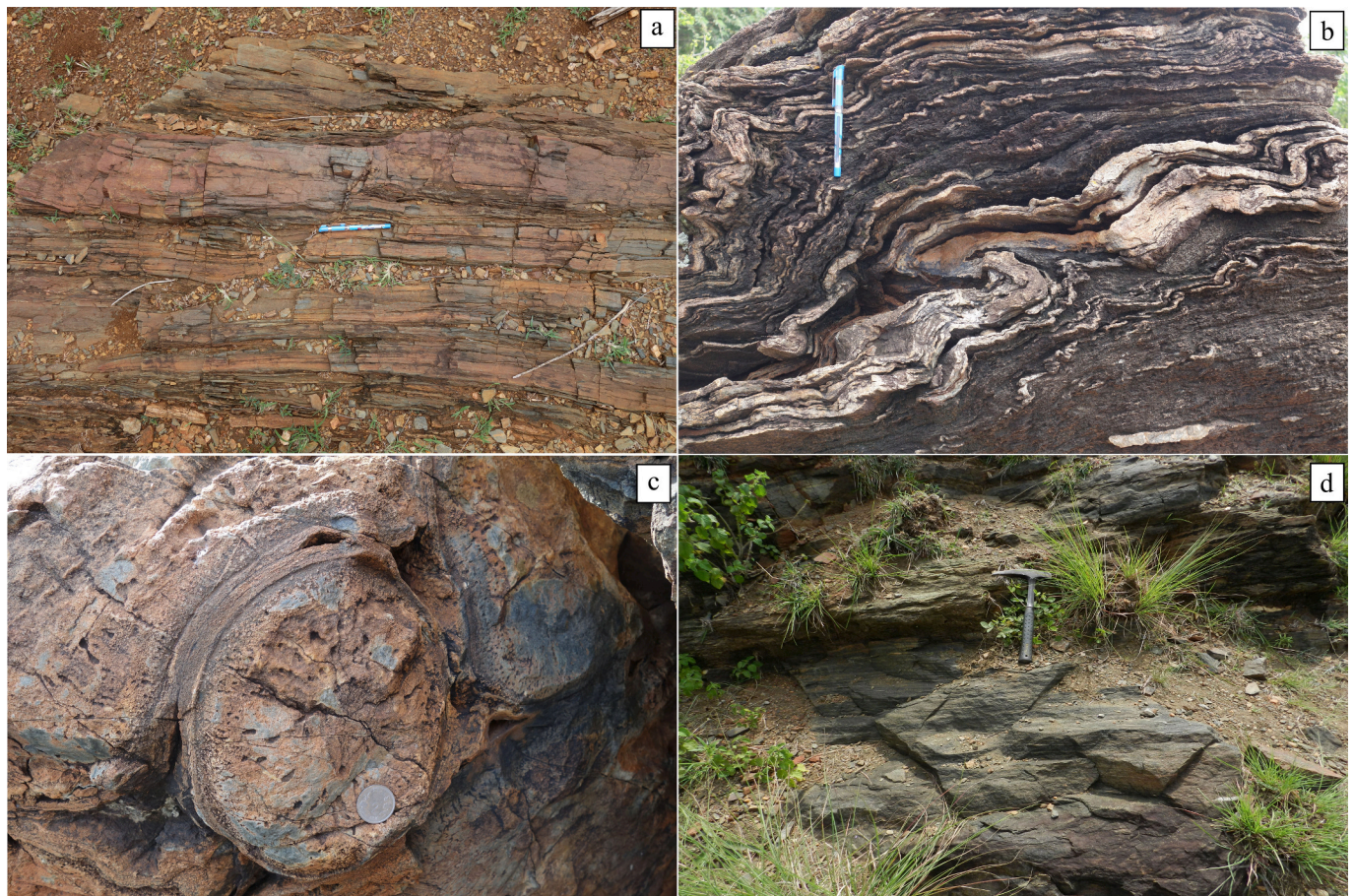


Fig. 7. (a) Mafic volcanics of Javagondanahalli belt interlayered with thin carbonate layers and BIFs (b) Tuff interlayered with carbonates (c) Spectacular pillow basalt in the northern outskirts of Maridihalli village (d) Amphibolite with shallow foliation in the Kunigal greenstone belt.

flows and rarely form crude pillows with chilled margins near Kallehadlu.

Voluminous greywacke -argillite -shale/carbonaceous shale, inter-bedded volcanic tuffs and BIFs corresponding to BIF III (Chadwick et al., 1981; Krapez et al., 2020) often containing sulphides (Figs. 3i & j) belong to the Hiriyur Formation found in the central to north-eastern part (Bellara-Ajjanahalli-Chikkasiddavvanahalli-Tekalvatti region). The observed lithological sequences quartzite-carbonate in the south-western part to greywacke-argillite-carbonaceous shales and sulphide bearing BIF III to the northeast imply transient conditions from shallower to deeper anoxic environments.

3.1.5. ca. 3000 Ma and 2600 Ma Potassic granite plutons

The most prominent potassic granite intrusions are ca. 3000-2950 Ma Bellur-Nagamangala pluton, Hosdurga granite (Sebastian et al., 2021, 2023; this study) and ca. 2600 Ma Arsikere - Banavara granite (Jayananda et al., 2006) which intrude the western basement TTG-type granitoids and Sargur Group ca. 3380 Ma old J.C. Pura greenstone belt whilst Chitradurga granite intrude Chitradurga greenstone belt (Jayananda et al., 2006). The Bellur-Nagamangala pluton contains quartz-monzonite, porphyritic monzogranite and pink granite which occasionally contains rounded to elongated mafic magmatic enclaves (Fig. 4a). The Hosdurga granite found as an NNW to SSE trending elongated pluton (see Fig. 1b) contains coarse grained light pink porphyritic monzogranite to coarse grained pink granite (Fig. 4b) and light grey granodiorite. The 2620 Ma Arsikere - Banavara plutons comprise highly coarse-grained light pink equigranular to porphyritic monzogranite, coarse grained dark grey quartz-monzogranite and medium grained pink granite (Jayananda et al., 2006). The 2614 Ma Chitradurga pluton (Fig. 4c) intrudes the volcanic-sedimentary sequences and contains voluminous coarse grained to porphyritic monzogranite that contain large K-feldspar phenocrysts and hornblende crystals (Fig. 4d) with minor dark grey granodiorite and light grey to pink granite.

3.1.6. ca. 3000 Ma Bukkapatna granite

The 2988±9 Ma old Bukkapatna granite (Chardon et al., 2011) runs nearly 50 km north-south as thin linear granite body bounded by the Chitradurga greenstone belt in the west and Javagondanahalli greenstone belt to the east (see Fig. 1b). The granite comprises voluminous light grey porphyritic monzogranite and subordinate grey to whitish grey granite. The granite is affected by strong deformation linked to regional shortening associated with shear deformation. It displays vertical to steep foliation with horizontal to shallow lineation. The Bukkapatna granite preserves all different stages of progressive shear deformation from C-S to pervasive C- prime fabrics, sinistral shear bands, eventually mylonite to ultra-mylonite (Chardon et al., 2011 and present study; Fig. 4e-f). The contacts between Bukkapatna granite and adjacent Chitradurga belt and Javagondanahalli belt are tectonized.

3.2. Chitradurga boundary shear zone

The Chitradurga boundary shear zone is a major high strain zone in the Dharwar craton that forms a dividing line between two crustal blocks (Swami Nath et al., 1976; Drury, 1983; Chadwick et al., 1989; Jayananda et al., 2006; Chardon et al., 2008, 2011). This shear zone is considered as a major crustal discontinuity and suture (Naqvi, 1973). The boundary shear zone comprises a network of large shear zones that traverse along the Chitradurga greenstone belt over 450 kms. Drury (1983) based on LANDSAT image analysis attributed Chitradurga boundary shear zone to a west-verging, east dipping thrust forming a major suture along which the western and eastern block were juxtaposed. Chadwick et al. (1989) proposed that the Chitradurga boundary shear zone represents a sinistral-sense transpressional regime rather than a thrust fault.

The boundary shear zone traverses Bukkapatna granite and

Javagondanahalli volcanics. The Bukkapatna granite show strain gradient with less deformed facies through C-S fabrics to C-primes with progressive grain size reduction and recrystallization leading to ultra-mylonite (see Fig. 4e-f). The adjacent Javagondanahalli greenstone volcanics to the east exhibit steep fabrics with reduction in the grain size and recrystallization. A progressive change in strain of the Javagondanahalli greenstone assemblages from less deformed shallow fabrics (Fig. 5a) near Sira through folded domain to steep mylonitic facies with recrystallization. West of Bukkapatna granite, the western branch of the boundary shear zone contains high-pressure assemblage (mica schist) at the intersection of NW-NE trending fabrics (Fig. 5b).

In the southern part of the boundary shear zone, large rotated boudins of melted garnet-amphibolite (Fig. 5c) found in the TTG-type migmatitic granitoid probably transported from deeper levels corresponding to mantle depth. Garnet-amphibolites show different stages of melting from the initiation of melting to accumulation of melt into thin leucosomes. Garnet is mainly confined to leucocratic melt fraction and is also found in unmolten residue. West of the boundary shear zone, the Chitradurga greenstone belt is traversed by anastomosing network of northward converging major dextral and sinistral shear zones at the latitude of 13° 30' (Chardon et al., 2008, 2011). Along these shear zones contain high pressure metamorphic assemblages like garnet-mica, - (paragonite?) -garnet- amphibole-clinopyroxene- found as large disrupted fragments probably transported from mantle depth (Fig. 5d).

3.3. Eastern corridor

3.3.1. Eastern basement: TTGs and transitional TTGs

Granitoids of TTG affinity and dark grey banded transitional TTGs (Sira gneiss complex; Chardon et al., 2011; Jayananda et al., 2020) with occasional dismembered outcrops of high grade supracrustal rocks (quartzite-pelite-calc-silicate-BIFs) with remnants of mafic rocks constitute the Eastern basement. The TTG-type granitoids are migmatitic to weakly banded gneisses occur as large discontinuous outcrops or enclaves within dark grey banded transitional TTGs. The TTG-type granitoids comprise dark grey to grey tonalite grey granodiorite and whitish grey trondhjemite which show shallow foliation (Fig. 6a) These migmatitic gneisses are traversed by melt filled NE trending ductile shear zones where the flat fabrics are transposed into steeper shear zones (Fig. 6b). These shear zones contain large dismembered boudins of migmatitic garnet-bearing amphibolite s (see Fig. 5c).

The transitional TTGs constitute voluminous lithologies of the eastern basement in the area between the Javagondanahalli belt and the Closepet batholith (see Fig. 1b) where they show spectacular flat to shallow fabrics (Fig. 6c & d). The transitional TTGs comprise abundant dark grey tonalitic to quartz-monzonite facies with minor grey granodiorite. Trondhjemite component typical of Archean TTGs is absent in transitional TTGs. These transitional TTGs contain numerous angular to ellipsoidal mafic boudins and occasional old migmatitic gneisses to mafic gneiss enclaves which are traversed by tiny leucocratic veins containing clinopyroxene and fluorite (Fig. 6e). They contain boudinaged tiny pink granite veins along foliation. The flat foliation of transitional TTGs show folding close to Javagondanahalli belt as well as along contact zone of the Closepet batholith. Several fine-grained whitish grey to light grey aplitic veins cutting across or found along the flat fabrics in the central part (e.g. Agrahara outcrop) which can be followed into the 2507±5 Ma old Maddakkanahalli pluton along the eastern boundary of the Javagondanahalli greenstone belt. The flat foliation progressively transposes into steep fabrics along the major NE trending shear zone where fluid flow has caused local melting and growth of large clinopyroxene, hornblende and fluorite crystals (e.g., Ravadi quarries, see Fig. 6e). South-west of Sira transitional TTGs contain thin whitish grey aplitic veins of Maddakkanahalli pluton along flat to shallow foliation (Fig. 6f).

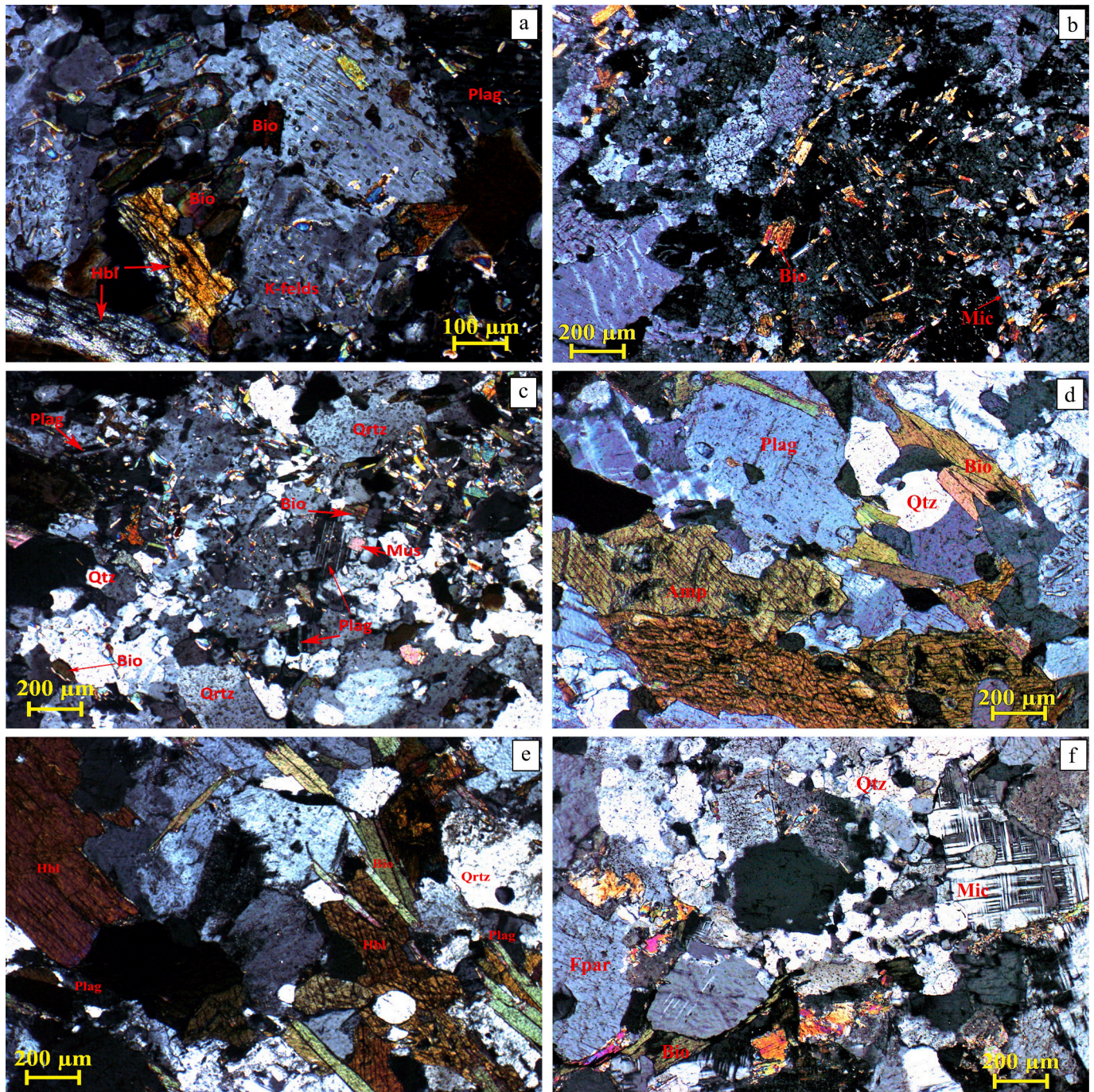


Fig. 8. (a) Western tonalitic gneiss showing medium to fine grained texture with sodic plagioclase, biotite, hornblende (b) sodic plagioclase with minor microcline, biotite, hornblende in granodiorite facies (c) plagioclase, biotite, hornblende with rare muscovite in trondhjemite (d) medium to coarse grained transitional TTG showing quartz, plagioclase, biotite and hornblende (e) coarse grained quartz monzonite displaying quartz, plagioclase, microcline, biotite, hornblende (f) fine grained aplitic Maddakkanahalli granite contain quartz, microcline, sodic plagioclase, minor biotite.

3.3.2. Javagondanahalli greenstone belt

The Javagondanahalli greenstone belt occurs as a curvilinear belt east of the boundary shear zone which extends from Nittur in the south up to Jagalur town to the north (see Fig. 1b). It is separated from the Chitradurga greenstone belt by ca. 3000 Ma Bukkapatna granite in the south and polymict Aimangala conglomerate in the northern part. Recent U-Pb zircon dating revealed that ca. 3000 Ma Bukkapatna granite extends up to the latitude of Chitradurga town (Mohan et al., 2023) which argues for the extension of Javagondanahalli greenstone belt further north of Chitradurga town. Javagondanahalli greenstone belt is bounded by large remnants of migmatitic TTGs to transitional TTG and

fragments of high grade supracrustal rocks to the east. Earlier studies (e.g., Swami Nath and Ramakrishnan, 1981) considered Javagondanahalli belt older than Chitradurga belt based on lithological associations and metamorphic grade (e.g., Swami Nath and Ramakrishnan, 1981). The Javagondanahalli greenstone belt comprises abundant metamorphosed mafic volcanics (Fig. 7a), tuffs with interlayered thin carbonate layers (Fig. 7b) and BIFs. Rarely mafic rocks preserve deformed pillows in the central part whilst spectacular pillows crop out in the northern part at Maridi Halli village (Fig. 7c). A gradation in mineralogy can be observed in mafic rocks from hornblende-plagioclase dominated amphibolite in the east to actinolite-epidote-plagioclase bearing facies, corresponding

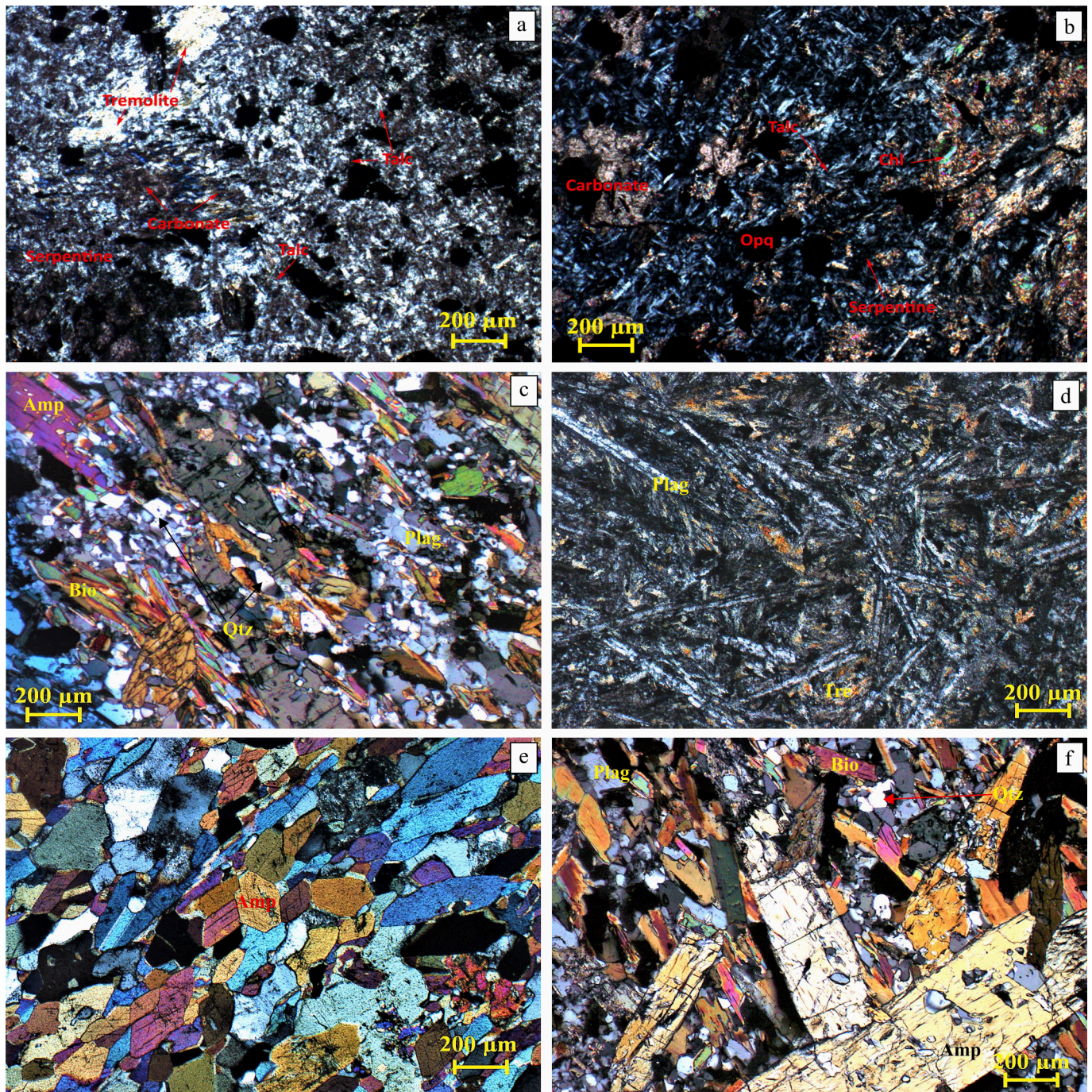


Fig. 9. (a) komatiite shows serpentine (rare olivine remnant), talc, tremolite (b) komatiite with serpentine with minor talc-chlorite (c) and (d) medium grained Chitradurga mafic volcanic show plagioclase, amphibole, actinolite, epidote with minor chlorite (e) Javagondanahalli amphibolite with plagioclase, hornblende (f) recrystallized Javagondanahalli tuff consist of quartz, plagioclase, actinolite, hornblende, and opaque phases.

to a decrease in metamorphic grade, towards the west close to the boundary shear zone. BIFs are confined to higher levels, forming summits and show alternate bands of iron oxides and chert.

Our detailed field observations reveal that the Javagondanahalli greenstone sequences extends to the north towards Jagalur. In this region the Javagondanahalli greenstone sequences are separated from Chitradurga greenstone belt with polymict Aimangala conglomerate whilst further north, it is separated from Chitradurga greenstone belt by a mylonitic boundary shear zone. East to northeast of Aimangala conglomerate, the Javagondanahalli greenstone sequences comprises metabasalts which often form spectacular pillows (see Fig. 7c), carbonate veins and BIF layers. Further north in G.R. Halli – Gonur region

metabasalts are traversed by NE to North trending quartz-carbonate veins.

The Javagondanahalli greenstone belt in its southern part (e.g., Sirabukkapatna section) exhibits a progressive change in fabrics from flat to folded domain to steep mylonitic fabrics to the boundary shear zone corresponding to an increase in strain from east to west. In the northern part, immediately east of the boundary shear zone, the metabasalts are traversed by two sets of cleavages (S1 and S2) which are often intersecting. The early cleavage (S1) probably corresponds to the emplacement of ca. 2.6 Ga Chitradurga granite and Jampalnaikankote granite (east of the boundary shear zone this granite intrudes into Javagondanahalli volcanics) whereas second cleavage (S2) appears to be

regional and coincides with development of a regional boundary shear zone (Jayananda et al., 2006; Chardon et al., 2008). At places across the boundary shear zone, S1 transposes into S2 cleavage.

3.3.3. Kunigal greenstone belt

The Kunigal greenstone belt is located 80 km south of Sira. It is a curvilinear 30 km N-S trending (see Fig. 1b) volcano-sedimentary belt (Swami Nath and Ramakrishnan, 1981) displays lithological associations similar to Javagondanahalli greenstone belt and possibly corresponds to its southern extension. These two greenstone sequences were displaced by a major sinistral shear zone (see Fig. 1b). The Kunigal greenstone belt is bounded by an older basement in the west and a mix of older basement, transitional TTGs and Closepet batholith to the east. The western basement shows progressive increase in the intensity of strain that eventually leads to a thick mylonitic Chitradurga boundary shear zone.

Along the eastern margin of Kunigal belt, discontinuous outcrops of high-grade metasedimentary units like quartzite, pelite, calc-silicate and BIFs found within the basement. On the contrary the western boundary of the Kunigal greenstone is bounded by boundary shear zone containing of TTG-type migmatitic gneisses contain large disrupted boudins of garnet-bearing melting amphibolites (see Fig. 5d).

The Kunigal greenstone belt contains abundant amphibolites with shallow foliation (Fig. 7d) in the east whilst foliation becomes progressively steeper to the west. Thin disrupted layers of ultramafic rocks are found in association with amphibolites in the western part. The sedimentary sequences found overlying amphibolite include thin layers of chert-carbonate - phyllites and BIFs to the summits.

3.3.4. Late (2550–2530 Ma) granitoid intrusions in the Eastern basement

Major granitoid bodies crop out east of Javagondanahalli belt which includes ~500 km long North-South trending Closepet batholith and its

offshoots into the adjoining basement (Jayananda et al., 1995, 2000; Jayananda et al., 2008; Jayananda et al., 2020; Gireesh et al., 2012). Few small leucocratic granite sheets and veins traverse the migmatitic basement along flat foliation of banded transitional TTGs (see Fig. 6d). Fine grained aplitic Maddakkanahalli granite pluton (Chardon et al., 2011) intrudes into the basement gneisses (migmatitic TTGs and banded transitional TTGs) and Javagondanahalli greenstone belt. This granite body is a highly homogenous light grey aplitic granite and crops out as large sheets along the flat foliation of basement gneisses (see Fig. 6f). Coarse-grained pegmatite dykes traverse the Maddakkanahalli granite which shows magmatic textures and do not display any visible fabrics on the outcrop scale.

The 2545–2520 Ma old Closepet batholith is a major geologic feature of the Dharwar craton and it forms the eastern limit of the studied corridor (see Fig. 1b). Regional fabrics analysis around the batholith reveals pervasive planar and linear fabrics with shallow east dipping foliation that is overprinted, in places, by steep foliations (Chardon and Jayananda, 2008). Large tracts of ca. 2600–2530 Ma juvenile magmatic rocks show flat to shallow foliations and extend from Javagondanahalli greenstone belt to Closepet batholith at the latitude of Sira (Chardon et al., 2011; Jayananda et al., 2015; Jayananda et al., 2020). This batholith is a composite magmatic complex that contains a dominant juvenile facies of sanukitoid affinity and a minor anatectic granite which originated from remelting of pre-existing crust (Jayananda et al., 2020).

3.4. Structural patterns

The studied E-W corridor represents a large crustal panel that displays diverse structures. Several structural studies (Naqvi, 1973; Drury and Holt, 1980; Drury, 1983; Mukhopadhyay et al., 1981; Chadwick et al., 1989, 2000, 2007) and strain fabrics analysis (Chardon et al., 1996, 2008, 2011; Choukroune et al., 1995; Jayananda et al., 2006)

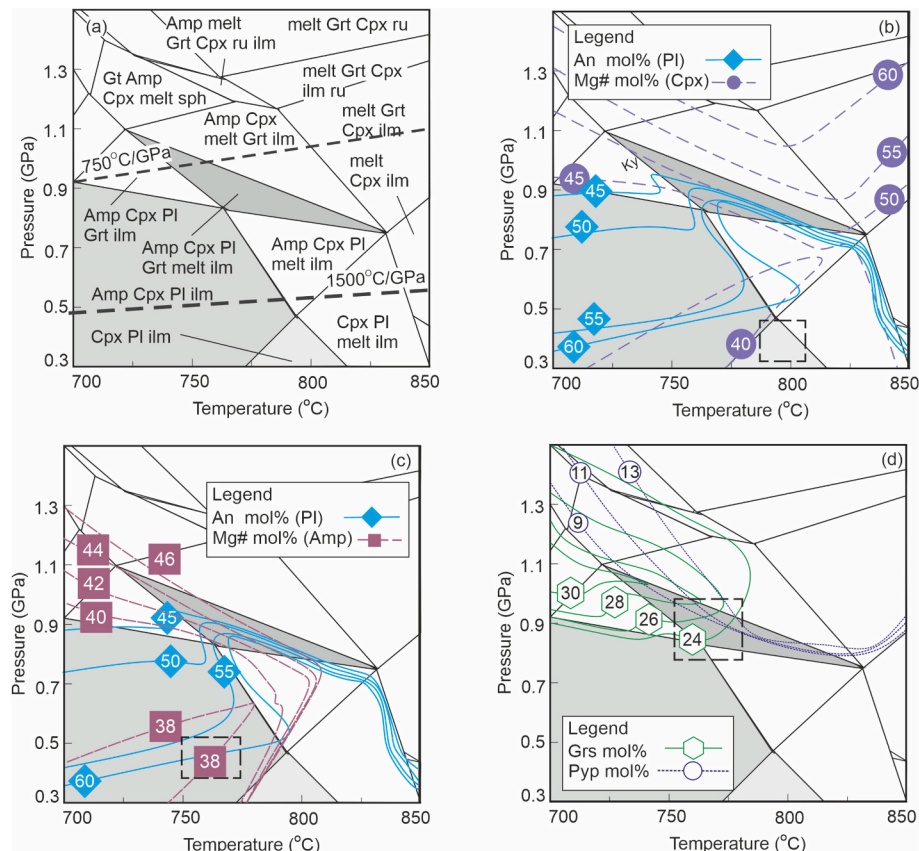


Fig. 10. (a-d) Metamorphic pseudosection and P-T diagram of (Al-01B) melting amphibolite along the western margin of Kunigal greenstone belt.

reveal dome and keel structures, regional scale north trending steep foliations, flat fabrics and regional strike-slip interference shear zone patterns. Chardon et al. (2008a, 2008b) have shown that craton scale shear zone pattern developed by interference between two shear zones; the NE trending dextral shear system in the south and the sinistral WNW trending shear system in the north. The central domain in the area between Chitradurga belt and Closepet batholith is a buffer zone where composite flat fabrics accommodating crustal flow are best preserved (Chardon et al., 2011).

The structure of the Chitradurga greenstone belt and adjoining basement is generally attributed to the fold-interference pattern (Mukhopadhyay et al., 1981; Naha et al., 1995). Chadwick et al. (1981) interpreted the fold interference pattern and steeply plunging antiformal structure around Chitradurga town due to single continuous deformation causing greenstone subsidence and uplift of the basement. LANDSAT image analysis reveals that regional shear zones traverse the Chitradurga greenstone belt and a major shear zone along the eastern margin is interpreted as marking west-vergent east dipping thrust along which western and eastern blocks amalgamated (Drury, 1983). Subsequently Chadwick et al. (1989) based on detailed tectonic fabrics data preclude Chitradurga boundary shear zone as a thrust but argued for a left-lateral transpressional shear zone. Detailed field-based fabrics data in combination with LANDSAT-TM data revealed asymmetric anastomosing network of sinistral shear zones displaying duplex like strike-slip structures (Jayananda et al., 2006). Furthermore, Jayananda et al. (2006) have documented two generations of fabrics in the Chitradurga greenstone belt and adjoining basement to the east. The greenstone volcanics adjacent mylonitic shear zone displays systematic superimposed fabrics suggesting an earlier deformation event in the greenstone belt which coincides with the emplacement of ca. 2615 Ma

Chitradurga granite. The late north to northwest trending steep fabrics with shallow lineation are compatible with regional shear zones traversing the basement developed close to 2500 Ma (Jayananda et al., 2006).

West of the Chitradurga greenstone belt, the basement granitoids form circular to elliptical gneiss domes whilst intervening greenstone sequences define keels in the J.C. Pura, Tiptur and base of Kibbanahalli arm (Chardon et al., 1996, 2008). These dome-keel structures are bounded by regional North – South trending transcurrent shear zones along their margins where both granitoid basement and greenstone sequences show steep fabrics. The Chitradurga boundary shear zone is a major regional high-strain zone that separates the western and central crustal blocks. The region comprises a network of NNW to NNE trending regional anastomosing sinistral shear zones displaying several duplex-like structures (Jayananda et al., 2006). These sinistral shear zones contain high-pressure assemblages including mica-rich metasedimentary rocks and garnet-clinopyroxene-amphibole bearing ultramafic rocks.

4. Petrographic characteristics

We present petrographic characteristics of different types granitoids across the Chitradurga boundary shear zone together with greenstone volcano-sedimentary sequences of Kibbanahalli, Chitradurga and Jagandanaahalli greenstone belts.

The western TTG-type granitoids are medium to fine grained and contain quartz, sodic plagioclase (An₁₂₋₂₀), minor microcline, biotite, hornblende in tonalite facies (Fig. 8a) and granodiorite (Fig. 8b), quartz, plagioclase (An₁₂₋₁₅), biotite with rare muscovite in trondhjemite (Fig. 8c). The common accessory minerals include zircon, apatite,

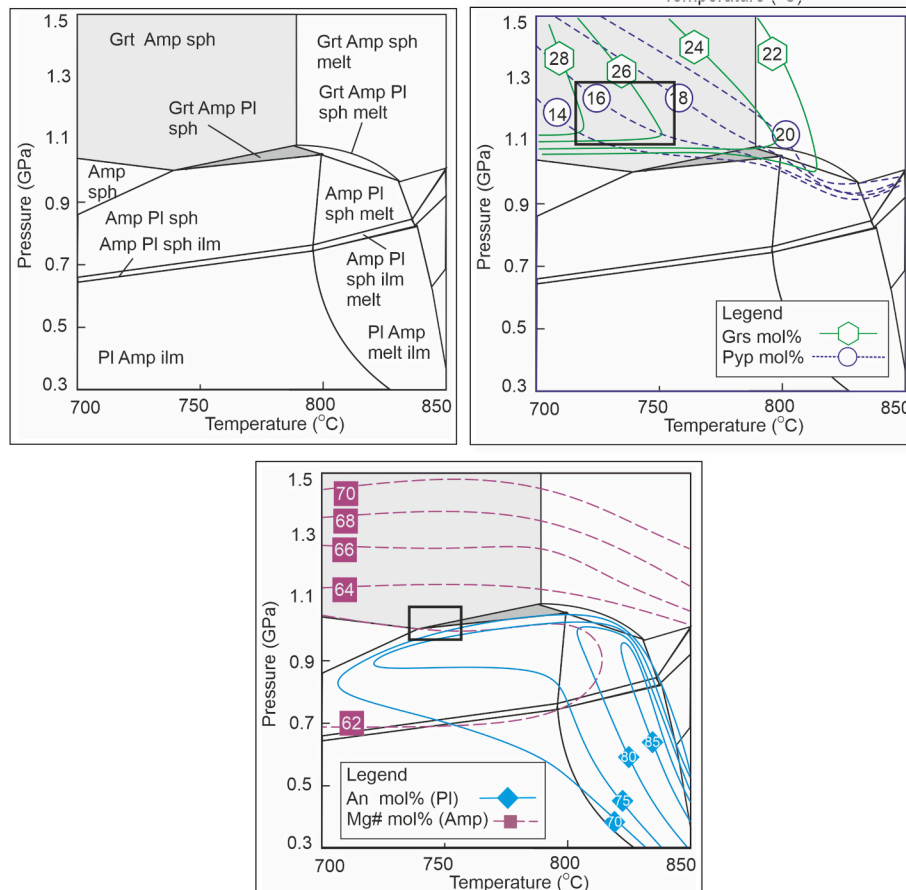


Fig. 11. (a-c) Metamorphic pseudosection and P-T diagram of (19SR-19) retrograded eclogite from interfering shear zone pattern NE of Chitradurga town.

titanite and opaques found as interstitial phases. The ca. 3000 Ma Hosdurga granite show medium to coarse grained equigranular to porphyritic texture. They consist of quartz, microcline, plagioclase (An₁₂₋₁₅), biotite, rare magmatic epidote, hornblende, muscovite with accessory zircon, apatite, allanite, titanite and opaque minerals.

In the eastern granitoid basement, the ca. 3300-3170 Ma migmatitic gneisses are medium to fine grained and contain quartz, plagioclase (An₁₀₋₂₃), minor microcline, biotite, occasional hornblende with accessory zircon, apatite and opaque phases (Fig 8d.). On the contrary the transitional TTGs are medium to coarse grained and have quartz, plagioclase (An₁₈₋₃₂) and microcline, biotite, hornblende, occasional clinopyroxene (Fig. 8e) with accessory phases like zircon, apatite, titanite and opaques in the mineral assemblage. The 2560-2540 Ma granitoids are composite and contains two distinct assemblages. The coarse grained dark grey quartz monzonite contains quartz, plagioclase (An₂₃₋₃₀), microcline, biotite, hornblende, occasional clinopyroxene together with accessory zircon, apatite, allanite, titanite and opaques, whilst light grey to pink granite contains abundant quartz, microcline, plagioclase (An₁₀₋₁₂), biotite with zircon, titanite and apatite as common accessory phases (Jayananda et al., 2020). The 2507 Ma aplitic Mad-dakkanahalli granite exhibit fine grained hypidiomorphic texture and contain quartz, microcline, sodic plagioclase (An₉₋₁₂), minor biotite with rare zircon and titanite (Fig. 8f).

The Sargur Group Ghattihsahalli, J.C. Pura, Banasandra and Nagamangala greenstone sequences contains abundant komatiites and komatiite basalts with minor basalts. Primary mineralogy is nearly absent except rare remnants of olivine. The observed mineralogy developed during post-magmatic hydrothermal alteration or low-grade metamorphism. The common mineral assemblages in komatiites include serpentine (rare olivine remnants), talc, tremolite (Fig. 9a) with occasional minor chlorite whilst komatiite basalts contain tremolite, talc, actinolite and chlorite (Fig. 9b) with or without minor serpentine. The basaltic volcanics exhibit actinolite, chlorite, rare tremolite and minor plagioclase (An₇₄₋₃₅) in the mineral assemblage.

Among the younger Dharwar Supergroup greenstone volcanics, the Kibbanahalli arm metabasalts are medium to fine grained basalts displaying vesicular to amygdaloidal structures. Primary mineralogy is absent, and the common mineral assemblage include plagioclase (An₃₅₋₄₆), actinolite, hornblende, occasional accessory calcite and quartz. The Chitradurga belt volcanics include fine grained metabasalts to fine to

medium grained felsic volcanics (pyroclastic flows). The common mineral assemblage of metabasalts include chlorite, actinolite, epidote, plagioclase (An₂₈₋₃₉) and opaque phases (Fig. 9c & d). Felsic volcanics are fine to medium grained with visible quartz crystals, aggregate of fine-grained quartz and feldspar, rare biotite with accessory zircon and opaques (Jayananda et al., 2013b).

The Javagondanahalli volcanics include fine to medium grained amphibolites and volcanic tuffs. The amphibolite comprises plagioclase (An₃₀₋₃₅), hornblende with or without actinolite and traces of quartz (Fig. 9e). The tuffs are recrystallized which consist of plagioclase (An₂₃₋₂₈), actinolite, hornblende, quartz and opaque phases (Fig. 9f). The Kunigal greenstone volcanics are mainly fine to medium grained schistose amphibolite consist of plagioclase (An₃₂₋₃₆), hornblende, minor actinolite, rare quartz and opaque phases.

5. Metamorphic record

Timing and P-T conditions of metamorphism in the studied corridor across the Chitradurga boundary shear zone and its spatial link to accretion, reworking and tectonic episodes in the Dharwar craton remains poorly constrained. Raase et al. (1986) in their study of the mafic rocks from Chitradurga greenstone belt documented a progressive increase in metamorphic grade from North to South. On the other hand, Rollinson et al. (1981) have described medium to high pressure metamorphic assemblages to the west of the Chitradurga greenstone belt whilst low pressure but high temperature assemblages to the east. Subsequent studies have shown three tectono-thermal events under moderate to high-pressure-temperature conditions (ca. 3100 Ma, 2950 Ma and 2500 Ma) to the west of the Chitradurga belt (Jayananda et al., 2013a; Dasgupta et al., 2019; Kaempf et al., 2025) and three metamorphic events to the east of boundary shear zone including ca. 3140 Ma amphibolite facies, 2640 Ma ultra-high temperature (UHT) event in the pelitic assemblages along the eastern margin of the Closepet batholith and a 2500 Ma amphibolite to granulite facies event affected the entire Archean crust of the Dharwar craton (Jayananda et al., 2011, 2013b; Kaempf et al., 2024). These thermal events spatially associated with juvenile magmatic accretion and crustal reworking leading to cratonization of the Archean crust (Chardon et al., 2011; Dey et al., 2016; Jayananda et al., 2006, 2020). It is crucial to understand the timing and P-T conditions of metamorphic events across the boundary shear zone as they bear on understanding tectonics, Neoproterozoic crustal evolution, and assembly of distinct crustal units and final amalgamation of crustal blocks into cratonic framework. Along the western margin of the Chitradurga greenstone belt quartz-muscovite-kyanite bearing pelitic sediments are found as large disrupted fragments interlayered with basement to the SW of Chitradurga at Kudineerkatte and further about 100 km south at Nelligere. Along the eastern boundary shear zone low pressure assemblages (cordierite-anthophyllite) are found as disrupted outcrops in the basement (Narayana et al., 1983). In this study, we present metamorphic P-T estimates for three samples from the Chitradurga boundary shear zone system which include a melted garnet bearing amphibolite (AL01B), an enclave from migmatitic TTG in the southern part of study, a mica schist (19 CH-01) in the central part of the boundary shear zone and a retrograded eclogite (19 SR19) from a NW trending shear zone traversing from transitional TTGs that merge into Chitradurga boundary shear zone to the north of Chitradurga town (see Fig. 1b)

5.1. Conventional thermobarometers

Empirical formulation (Holland and Blundy, 1994; Molina et al., 2015) involving co-existing compositions of amphibole and plagioclase is used to constrain the P-T condition for melt present (*AL-01B*), melt absent (*19SR 19*) retrograded eclogite, and mica schist (*19 CH-01*). However, for the two samples (Al-01B and 19SR-19), most commonly cited, the Fe-Mg exchange thermometer involving garnet and amphibole

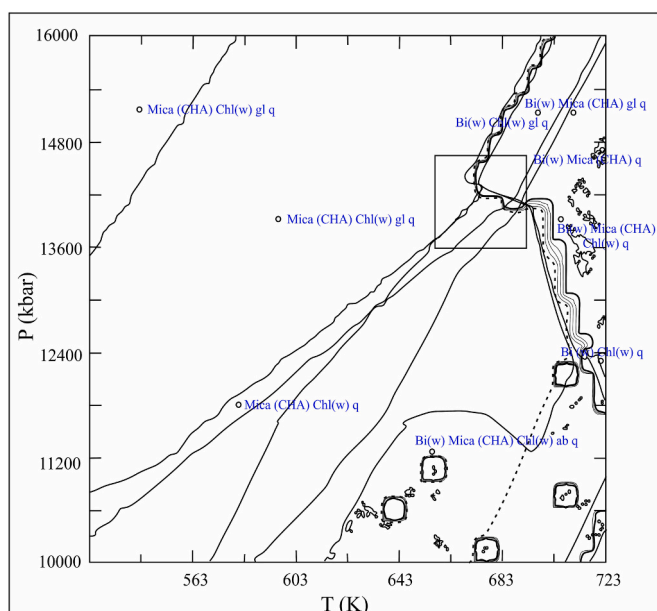


Fig. 12. Metamorphic P-T diagram of (19CH-01) sodic-mica from high-strain zone traversing the Chitradurga greenstone belt NW of Ajjanahalli mines.

is not used for temperature estimation due to an uncertainty in the $\text{Fe}^{3+}/\text{Fe}^{\text{total}}$ values computed from amphibole structural formulae.

5.2. P-T pseudosection analysis

The P-T pseudo-sections are to structure for the whole-rock chemical composition (TiNCKFMASH) by using PERPLEX (Connolly, 2009) to model phase stability and reaction paths for the studied samples. The thermodynamic properties of the minerals are taken from hp633ver.dat (Holland and Powell, 2011). The following activity-composition models are used: augite, amphibole, and silicate melt (Green et al., 2016), feldspar (Benisek et al., 2010), and garnet (White et al., 2007). The stoichiometry of MnO and P_2O_5 is adjusted with FeO_t and CaO , respectively. The SiO_2 and H_2O are set to excess, as the studied sample contains quartz and hydrous minerals like amphibole.

5.2.1. Sample AL-01B - melting of gt-amphibolite

The TiNCKFMASH P-T pseudosection constructed using the whole-rock bulk chemical composition for temperature 700 to 850°C and pressure 0.3 to 1.5 GPa is shown in Fig. 10a. In the P-T pseudosection, the plagioclase – clinopyroxene – silicate-melt – ilmenite is stable at temperature 800 to 850°C at pressure < 0.9 GPa. A decrease in temperature < 800°C leads to the disappearance of silicate melt, and plagioclase – clinopyroxene – ilmenite becomes a stable assemblage. Amphibole joins the plagioclase – clinopyroxene – ilmenite assemblage at temperature 775°C, and the assemblage, plagioclase – clinopyroxene – ilmenite – amphibole, become stable between temperature 700 to 775°C and pressure 0.3 to 0.9 GPa. As most of the amphibole contain relict clinopyroxene and plagioclase inclusions, the *amphibole-in* line marks the temperature limit the rock has experienced for stabilizing amphibole by consuming plagioclase and clinopyroxene.

An increase in pressure > 0.9 GPa, between 700 to 750°C, garnet appears to be a stable phase with plagioclase – clinopyroxene – ilmenite – amphibole assemblage. The nearly horizontal slope of the *garnet-in* univariant reaction line marks the minimum pressure the rock has experienced. Silicate melt is stable with plagioclase – clinopyroxene – ilmenite – amphibole – garnet at temperature > 750°C and pressure > 0.9 GPa. An increase in pressure > 0.9 to 1.1 GPa between 750 – 850°C leads to destabilization of plagioclase. In addition, amphibole destabilized between 800 – 850°C between 0.3–1.5 GPa. As the observed assemblage contains both plagioclase and amphibole with garnet and melt, the *plagioclase-out* and *amphibole-out* invariant reaction lines define the maximum pressure and temperature the rock has experienced.

The chemical composition of clinopyroxene, plagioclase, and amphibole is modelled and compared with the measured composition to constrain the P-T conditions. Isopleths corresponding to the measured composition of clinopyroxene and plagioclase, i.e., $\text{Mg\# mol\%} = 40$, and $\text{An mol\%} = 55 - 60$, intersect at 775 – 800°C, 0.3 - 0.5 GPa very close to the clinopyroxene – plagioclase – ilmenite field (Fig. 10b). The modelled $\text{Mg\# mol\%}$ of amphibole (38 – 40) intersect with the plagioclase feldspar's model composition (55 – 60) at a slightly lower temperature ~ 750°C, 0.3 - 0.5 GPa (Fig. 10c). The P-T estimates 775 – 800°C, and 75°C, obtained from the combination of clinopyroxene – plagioclase feldspar and amphibole – plagioclase feldspar composition, imply initial hydration and colling from magmatic conditions to amphibole bearing assemblages. The peak P-T condition, 750 – 800°C, 0.9- 1.1 GPa (Fig. 10d), is constrained from the intersection of pyrope (8-12) and grossular (24–28) composition, similar to those calculated from mineral-chemical data. Compared with the amphibole forming condition P-T conditions, the garnet peak P-T condition implies a prograde segment of a clockwise P-T path with an incongruent breakdown of amphibole producing garnet and silicate melt.

5.2.2. Sample SR19 – retrograded eclogite

The TiNCKFMASH P-T pseudosection constructed using the whole-rock bulk chemical composition for 700 to 850°C and 0.3 to 1.5 GPa

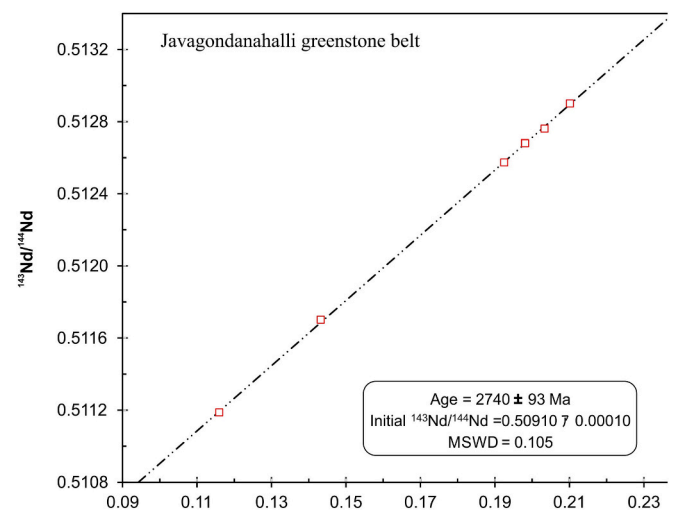


Fig. 13. Sm-Nd whole rock isochron of mafic volcanic rocks of Javagondanahalli greenstone belt.

is shown in Fig. 11a. In the phase diagram, plagioclase feldspar – amphibole – ilmenite is stable between 700 – 800°C, 0.5 – 0.6 GPa. Silicate melt appears as a stable phase with plagioclase feldspar – amphibole – ilmenite at > 800°C. Since the studied sample does not contain melt, the *melt-out* invariant line contains the maximum temperature the rock has experienced during the stabilization of plagioclase – amphibole – ilmenite assemblage. Increasing pressure leads to the appearance of sphene by consuming ilmenite. Still, at increased pressure > 1.1 GPa, the plagioclase absent, amphibole – garnet – sphene assemblage is stable. As the rock mainly consists of garnet and amphibole, the *garnet-in* invariant line constrains the minimum pressure the rock has experienced. The plagioclase-amphibole symplectite around garnet is marked by the plagioclase – amphibole – garnet assemblage between 750 – 800°C, at 1.0 GPa (see Fig. 11a).

Since garnet – amphibole is the peak P-T assemblages, the intersection of the chemical composition of pyrope (14-18) and grossular (26 – 28) at 700 – 750°C > 1.2 GPa constrain the peak P-T condition (Fig. 11b). Compared with the metamorphic phase diagram, the peak P-T condition coincides with the lower limit of eclogite facies assemblages. Accordingly, the symplectite texture, i.e., plagioclase-amphibole corona around garnet, indicate nearly isothermal decompression to < 1.0 GPa, from peak P > 1.2 GPa (Fig. 11c).

5.2.3. Sample 19 CH01

In sample 19 CH01 (Fig. 12), the P-T pseudosection is divided into three distinct fields based on the assemblages. The garnet bearing assemblage is stable at high P-T conditions. In contrast, the biotite bearing assemblage is stable at low P-high T conditions. Furthermore, the mica- chlorite assemblage is stable at relatively high P – low T metamorphic conditions. Since, the samples studied do not contain biotite or garnet. Since the sample is extremely fine grained, with sub-micron particles dispersed along the mica- chlorite aggregate, we suggest that the stability field of chlorite- biotite and other accessory phases define the max P-T conditions for the assemblage. In a P-T field, the 19 CH01- and SR-19 when plot a crustal segment where the 19 CH-01 exhumed from deeper part of colder thickened crust in the western part of boundary shear zone whilst SR-19 represent exhumed a deeper section of hotter crust close to the contact between two blocks.

6. Geochronology

In order to constrain the time frame of crustal accretion, reworking and the metamorphic record, we have carried out U-Pb zircon geochronology of the TTG basement and potassic granite plutons to the

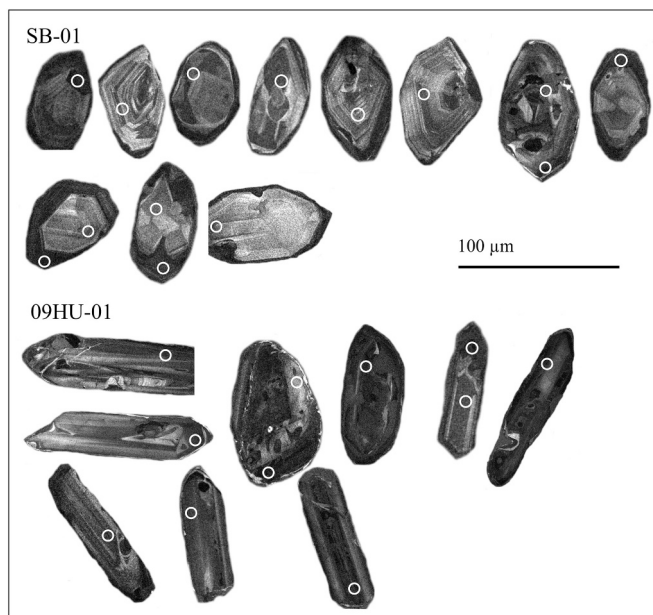


Fig. 14. Cathodoluminescence images of zircon grains from the TTG gneiss samples of SB-01 and 09HU-01 (white circle denotes the U-Pb analysed spot).

west of the Chitradurga greenstone belt as well as TTGs/transitional TTGs and granitoid intrusions to the east of the boundary shear zone. We also present Sm-Nd whole rock isochron for the mafic volcanics of Javagondanahalli greenstone belt. Published whole-rock isochrons, U-Pb zircon, monazite ages coupled with in-situ Lu-Hf dating of garnet are also discussed.

6.1. Western corridor

6.1.1. ca. 3400–3180 Ma TTGs

Earlier Rb-Sr isotope data define a whole isochron define age of 3028 ± 28 Ma for the weakly migmatitic gneisses from the western margin of Chitradurga belt (Taylor et al., 1984). U-Pb zircon ages revealed a major TTG forming event at ca. 3310–3209 Ma to west of the Chitradurga belt at the latitude of Chitradurga (this study; Chardon, 1997; Ravindran et al., 2023). More recently Mohan et al. (2023) have published U-Pb zircon ages of 3208 ± 11 Ma, 3217 ± 4 Ma, 3206 ± 5 Ma and 3213 ± 4 Ma for TTGs and U-Pb titanite ages of ca. 3039–3050 Ma. Titanite U-Pb ages for the same rocks which are in consistent with U-Pb zircon age of 3197 ± 11 Ma for the basement TTG gneiss. The titanite ages are different from those of zircon ages but match the later magmatic event that probably affected the surrounding granitoids. Therefore we do not interpret the age of titanites as magmatic but rather marking local metamorphism.

Further southwest in the Tiptur region (see Fig. 1b), zircons from TTG yield U-Pb ages of ca. 3200–3150 Ma (Chardon, 1997) which are also consistent with the U-Pb zircon ages of 3270–3100 Ma obtained further southwest with xenocrysts as old as 3400–3300 Ma (Ao et al., 2021; Wang et al., 2023). These published U-Pb zircon ages together with our new U-Pb zircon ages presented in this study (presented in the following section) reveal two major TTG-crust forming event including an early minor ca. 3400–3300 Ma and a major crust forming event during ca. 3230–3150 Ma. These two major TTG-crust forming events are spatially associated with komatiite dominated greenstone volcanism west of the Chitradurga greenstone belt in Ghattihosahalli, J.C. Pura and Banasandra belts (Jayananda et al., 2008, 2016, 2023; Maya et al., 2017).

6.1.2. ca. 3000 Ma potassic granites

The high potassic granite plutons in the Bellur-Nagamangala granite

in the southern region, Hosdurga granite, Bukkapatna granite in the northern part region and Arsikere granite in the central part of the corridor intrude the TTG basement and associate Sargur Group sequences. Zircon U-Pb ages of Bellur-Nagamangala granite yield ages of ca. 3000–2950 Ma (present study; Jayananda et al., 2019) which are consistent with a U-Pb zircon 3000 Ma for granite obtained by Ao et al. (2021) and Wang et al. (2023). Further north, zircons from Hosdurga granite yield U-Pb Concordia age 2954.8 ± 4.5 Ma (present study) which are in the range of U-Pb zircon ages 3010 ± 25 Ma and 3026 ± 26 Ma obtained for granite from Ghattihosahalli area (Ravindran et al., 2023). Zircons from mylonitic Bukkapatna granite yield U-Pb ages of 2988 ± 4 Ma (Chardon et al., 2011) and its northern extension close to Chitradurga town yield 2973 ± 6 Ma and 2962 ± 6 Ma respectively (Mohan et al., 2023).

6.1.3. ca. 2620–2610 Ma potassic granites

The Chitradurga granite pluton intrude the Chitradurga greenstone belt and adjacent basement, whilst Arsikere-Banavara granite and Pandavapura granite (Seshadri et al., 1981; Taylor et al., 1984; Jayananda et al., 2006, 2019; Chadwick et al., 2007; Mohan et al., 2014) intrude the basement. U-Pb zircon dating of these potassic granites indicate 2620–2590 Ma for their magmatic emplacement (Jayananda et al., 2006, 2019; Ram Mohan et al., 2014; Chadwick et al., 2007).

To summarize, the published U-Pb zircon ages reveal TTG-type granitoids of the western basement formed during two major episodes ca. 3400–3300 Ma and 3250–3150 Ma and potassic granite emplaced during ca. 3000–2950 Ma and ca. 2620–2600 Ma.

6.2. Greenstone volcanics

Published ages related to the timing of greenstone volcanism from the Sargur Group, Bababudan Group and Chitradurga Group are constrained by whole-rock Sm-Nd isotopes on ultramafic–mafic volcanics and U-Pb zircon dating of felsic volcanics.

6.2.1. ca. 3400–3150 Ma Sargur Group komatiites

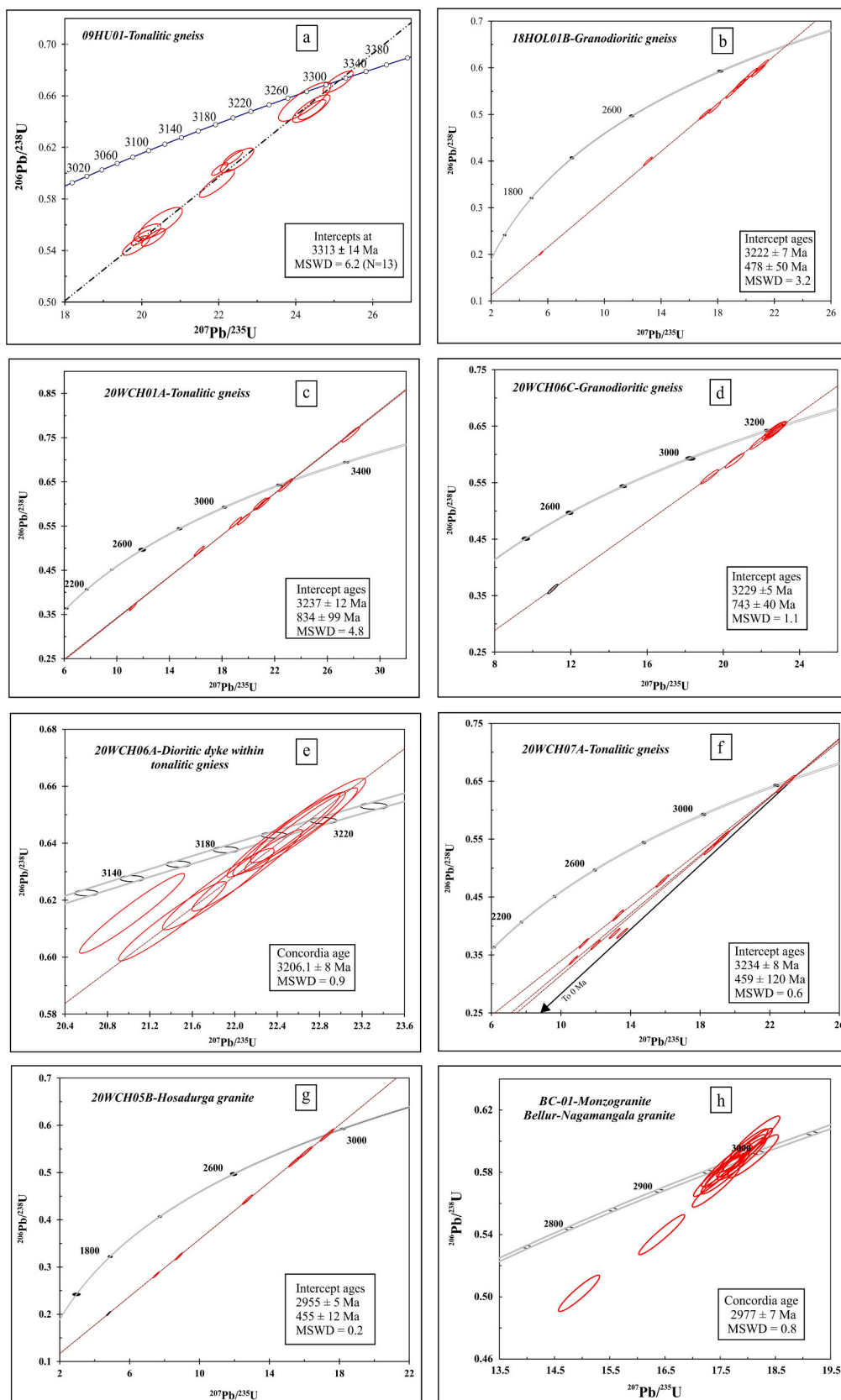
Jayananda et al. (2008) obtained an Sm-Nd whole-rock isochrone age of 3352 ± 110 Ma for komatiite lava flows from the Sargur Group greenstone sequences and 3384 ± 200 Ma for the J.C. Pura komatiite flows. More recent Sm-Nd whole-rock isochron for komatiite lava flows from the stratigraphically equivalent Sargur Group greenstone belts of Kalyadi, Nuggihalli, J.C. Pura and Holenarsipur greenstone belt provide an age of 3356 ± 68 Ma (Jayananda et al., 2023) whilst zircons from felsic pyroclastic flows from higher stratigraphic levels indicate a U-Pb age of 3305 ± 10 Ma (Jayananda et al., 2023), which is in agreement with U-Pb zircon age of 3298 ± 7 Ma for felsic volcanic rock (Peucat et al., 1995). Sm-Nd whole-rock isochrons yielded slightly younger ages of ca. 3284–3150 Ma for Kalyadi, Nuggihalli and Banasandra greenstone belts (Jayananda et al., 2008; Maya et al., 2017; Mukherjee et al., 2012). More recently Sargur Group metatuff from the western margin of the Chitradurga belt indicate U-Pb age of 3313 ± 6 Ma (Sreehari and Toyoshima, 2020).

6.2.2. ca. 2900–2850 Ma Bababudan Group mafic volcanics

Mafic volcanic rocks from lower stratigraphic levels (Kalasapura Formation) of the Bababudan basin yielded an Sm-Nd whole-rock isochron age of 2911 ± 49 Ma whilst higher stratigraphic levels (Santaveri Formation) yielded 2848 ± 70 Ma (Kumar et al., 1996). The Kibbanahalli arm is considered to be stratigraphically equivalent of Bababudan Group (Venkata Dasu et al., 1991) and the timing of eruption of volcanics probably close to ca. 2900–2800 Ma.

6.2.3. ca. 2740–2676 Ma Chitradurga Group mafic and felsic volcanics

Isotopic age data for the volcanic rocks corresponding to different stratigraphic levels of the Chitradurga Group are presented here. Kumar et al. (1996) published a Sm-Nd whole-rock age of 2747 ± 15 Ma for



(caption on next page)

Fig. 15. Zircon U-Pb concordia diagrams of granitoids to the west of boundary shear zone and 3.0 Ga potassic plutons (a) U-Pb concordia diagram of tonalitic gneiss (09 HU-01) from western margin of the Chitradurga greenstone belt close to Huliya (b) U-Pb concordia diagram of granodiorite (18 HOL-01B) from Holalkere (c) U-Pb concordia diagram of tonalite (20 WCH-01A) from Siranakatte dome (d) U-Pb concordia diagram of grey granodiorite (20WCH-06C) from Bharamanayakanadurga (e) zircon U-Pb concordia diagram of dioritic dyke within tonalitic gneiss (20WCH-06A) from Bharamanayakanadurga (f) U-Pb concordia diagram of tonalitic gneiss (20WCH-07A) from Holalkere-Chitradurga road (g) U-Pb zircon concordia diagram of Hosdurga potassic granite (20WCH-05B) near Janakal village (h) U-Pb zircon concordia diagram of grey monzogranite (BC-01) near Bellur cross.

volcanics of Chitradurga Group. However, their Sm-Nd whole-rock isochron included samples of metabasalts and pyroclastic flows from different stratigraphic levels. On the other hand, Nutman et al. (1996) provided SHRIMP U-Pb age of 2614 ± 8 Ma for the pyroclastic flows from Dagenikatte in the Shimoga basin that probably younger to Chitradurga Group. SIMS U-Pb zircon age 2677 ± 2 Ma was obtained for a pyroclastic flow from higher stratigraphic level (Jayananda et al., 2013a) coincides with SHRIMP U-Pb zircon age of 2676 ± 20 Ma for a rhyolite sill (Hokada et al., 2013) from the same region. Direct dating of low-grade metamorphic carbonate from Chitradurga belt using whole-rock Pb-Pb method defines an isochron age of 2639 ± 32 Ma (Russell et al., 1996).

6.2.4. Bukkapatna granite in the Chitradurga boundary shear zone

Zircon U-Pb dating of Bukkapatna granite in its type locality provided an age of 2988 ± 9 Ma (Chardon et al., 2011). More recent U-Pb dating of three samples of Bukkapatna granite further north close to Chitradurga town indicate age of 2973 ± 9 Ma, 2992 ± 6 Ma and 2962 ± 11 Ma (Mohan et al., 2023).

6.3. Eastern corridor - (ca. 3300-2500 Ma)

6.3.1. TTGs -Transitional TTGs and granitoid intrusions

Zircons from migmatitic TTG gneisses along the eastern margin of the Javagondanahalli belt indicate U-Pb concordia upper intercept ages of 3314 ± 38 Ma (Nasheeth et al., 2015) whilst further south a large migmatitic gneiss enclave in transitional TTG yield ages of 3231 ± 5 Ma, 3226 ± 8 Ma and a xenocryst indicate 3375 ± 13 Ma (Chardon et al., 2011). North of Challakere zircons from migmatitic gneisses yield two age groups ca. 3222-3271 Ma and 3301-3360 Ma (Jayananda et al., 2013b; Jayananda et al., 2020). On the contrary the transitional TTGs in the area between the Javagondanahalli belt and Closepet granite provide U-Pb zircon ages of 2669.9 ± 7.7 Ma, 2601 ± 16 Ma and 2600 ± 11 Ma (Jayananda et al., 2020). Juvenile magmatic intrusions within transitional TTGs provide zircon U-Pb ages of 2573 ± 8 Ma, 2568 ± 11 Ma and 2562 ± 4 Ma (Jayananda et al., 2020; Chardon et al., 2011). Leucocratic fine grained aplitic granite injects along flat foliations of gneisses yield U-Pb zircon age of 2507 ± 5 Ma (Chardon et al., 2011).

6.3.2. Javagondanahalli greenstone belt

The timing of volcanism in the Javagondanahalli greenstone belt is not precisely known although stereographically considered as older than Chitradurga belt (Swami Nath and Ramakrishnan, 1981). In order to constrain the timing of volcanism in the Javagondanahalli belt and its time relationship with the Chitradurga greenstone belt, we have performed Sm-Nd isotope analysis of whole-rock samples and data are presented in the Supplementary table 1. All the analysed samples are mafic to intermediate in composition and display large variations in Sm/Nd ratios. This is also reflected in their distinct REE patterns shown in Fig. 19. Six out of ten analysed whole-rock samples define an isochron age of 2740 ± 93 Ma (MSWD = 0.11 at $\pm 2\sigma$ level) with initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.50910 ± 0.00010 ($\pm 2\sigma$) (Fig. 13). The four discarded samples show slightly disturbed REE patterns probably during the fluid induced alteration processes associated with metamorphism close to 2500 Ma. Consequently, the six-point Sm-Nd whole-rock isochron age of 2740 ± 93 Ma is interpreted as eruption of mafic volcanics in the Javagondanahalli greenstone belt. This age is contemporaneous with published whole-rock Sm-Nd isochron age of 2747 Ma for

the mafic-felsic volcanics of the Chitradurga greenstone belt (Kumar et al., 1996). In summary, our new results together published Sm-Nd isochron reveal mafic volcanism in Javagondanahalli greenstone belt is contemporaneous with mafic volcanism in the Chitradurga belt dated ca. 2740 Ma.

6.4. New U-Pb zircon ages of the granitoids: Western corridor

6.4.1. Sample 09 HU-01 - 3313 ± 14 Ma

Sample 09 HU-01 is a medium to coarse grained tonalitic gneiss collected at Keresurgondanahalli village ($13^{\circ}35'02''$ N; $76^{\circ}30'02''$ E) in Huliya - Hosdurga road. Zircon grains are slightly elongated, euhedral to subhedral and display simple oscillatory zoning with few darker areas (Fig. 14). Zircon crystals have low to high U (38-1299 ppm) and Th (79-1021 ppm) respectively with corresponding Th/U ratios of 0.16 to 1.3. Most concordant analysis of 13 grains define a concordia upper-intercept age (Fig. 15a) of 3313 ± 14 Ma (MSWD = 2σ) with a lower-intercept age of 531 ± 250 Ma. This concordia upper-intercept age is interpreted as crystallization of magmatic precursors of the tonalitic gneiss. The obtained age of 3313 ± 14 Ma coincides with the age of oldest TTG basement further south-west in Tiptur region (Chardon, 1997).

6.4.2. Sample 18HOL-01B - 3222 ± 7 Ma

Sample 18HOL-01B is a coarse-grained granodiorite collected from large quarry at Jayanthi Nagar ($14^{\circ}01'57''$ N; $76^{\circ}08'59''$ E) about 5 km west of Holalkere town.

Zircon crystals imaged and analyzed from 18-HOL-01B all showed simple fine-oscillatory zoning from core to rim, though with some localized darker domains indicative of radiation-damage accumulation (Fig. 16). Similarly, U-Pb data distribution in concordia plot (Fig. 15b) is simple with an upper-intercept age of 3222 ± 7 Ma (2σ) and a lower-intercept at 478 ± 50 Ma. U-Pb data exhibit significant discordance between 6% to 63%. Zircons analyzed had low to moderately high U and Th concentrations of 221-729 and 114-425 ppm, respectively, associated with variable Th/U values of 0.31-1.08.

6.4.3. Sample 20 WCH-01A - 3237 ± 12 Ma

Sample 20 WCH-01A is a tonalite collected from Siranakatte dome in Hiriya-Hosdurga road ($13^{\circ}53'50''$ N; $76^{\circ}25'57''$ E).

Zircon crystals have very simple fine-oscillatory zoning from core to rim, though they appear darker in CL images than zircons from previously described samples (Fig. 16). U-Pb data form a single discordia line in a concordia plot (Fig. 15c) that indicates an upper-intercept age of 3237 ± 12 Ma and a lower-intercept one at 834 ± 99 Ma (2SE ; MSWD = 4.8). The upper-intercept age is completed by a concordia age of 3231 ± 11 Ma (2σ MSWD = 1.9), which we interpret as the igneous crystallization age of this sample. In terms of concentrations, zircons have low to moderate U and Th contents (120-627 and 56-251 ppm, respectively), and they are associated with a narrow range of Th/U values (0.30-0.60).

6.4.4. Sample 20 WCH-06C - 3229 ± 5 Ma

Sample 20 WCH-06C is a coarse to medium grained grey granodiorite collected from the abandoned quarry in the northern margin of Bharamanayakanadurga ($14^{\circ} 11'13''$ N; $76^{\circ} 07'43''$ E). Zircon crystals commonly euhedral to subhedral and show fine-oscillatory zoning from core to rim with moderate U and Th concentrations (234-432 and 81-192 ppm, respectively) and Th/U ratios ranging from 0.23 to 0.63

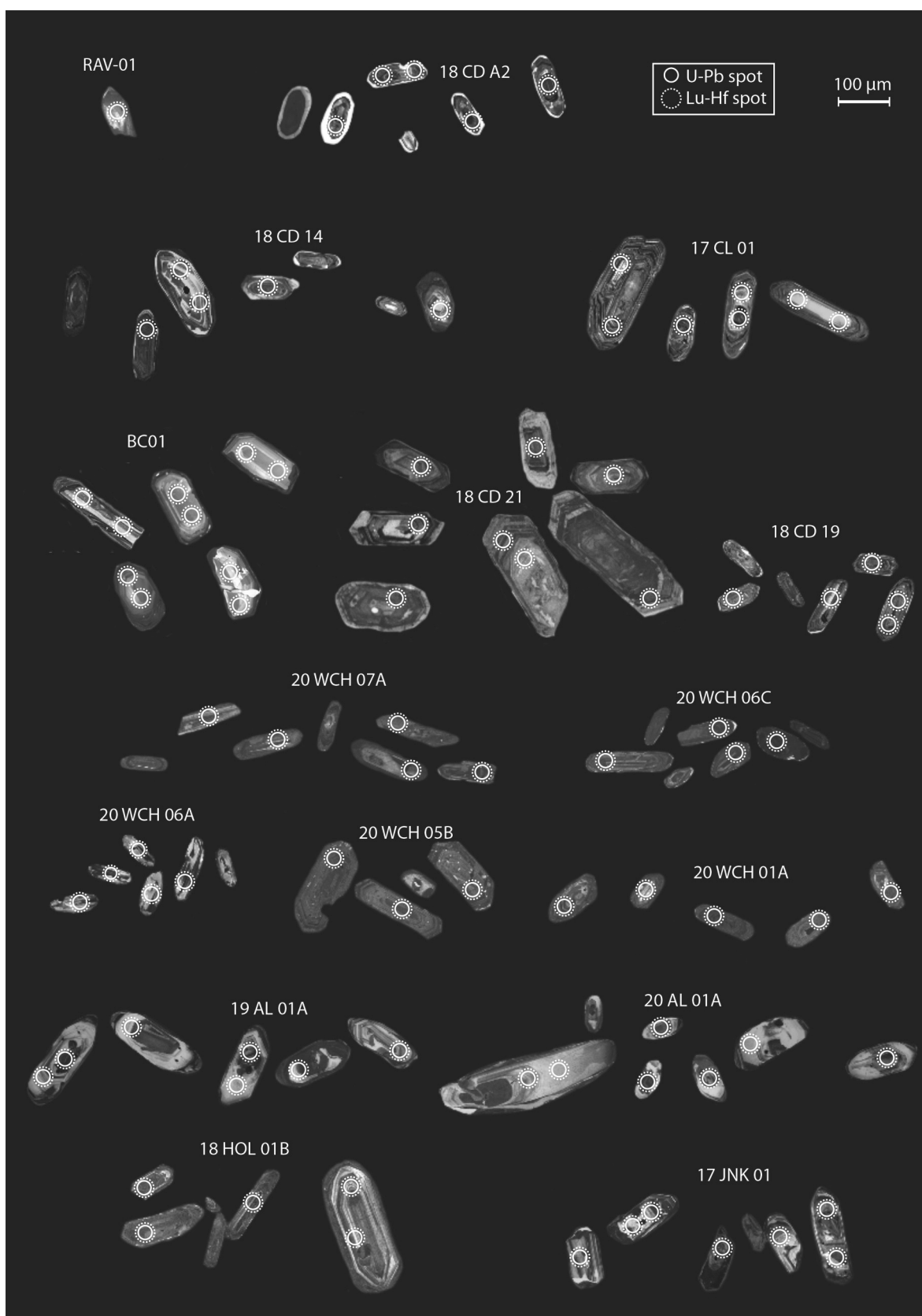


Fig. 16. Cathodoluminescence images of zircons from the western and eastern basement granitoids.

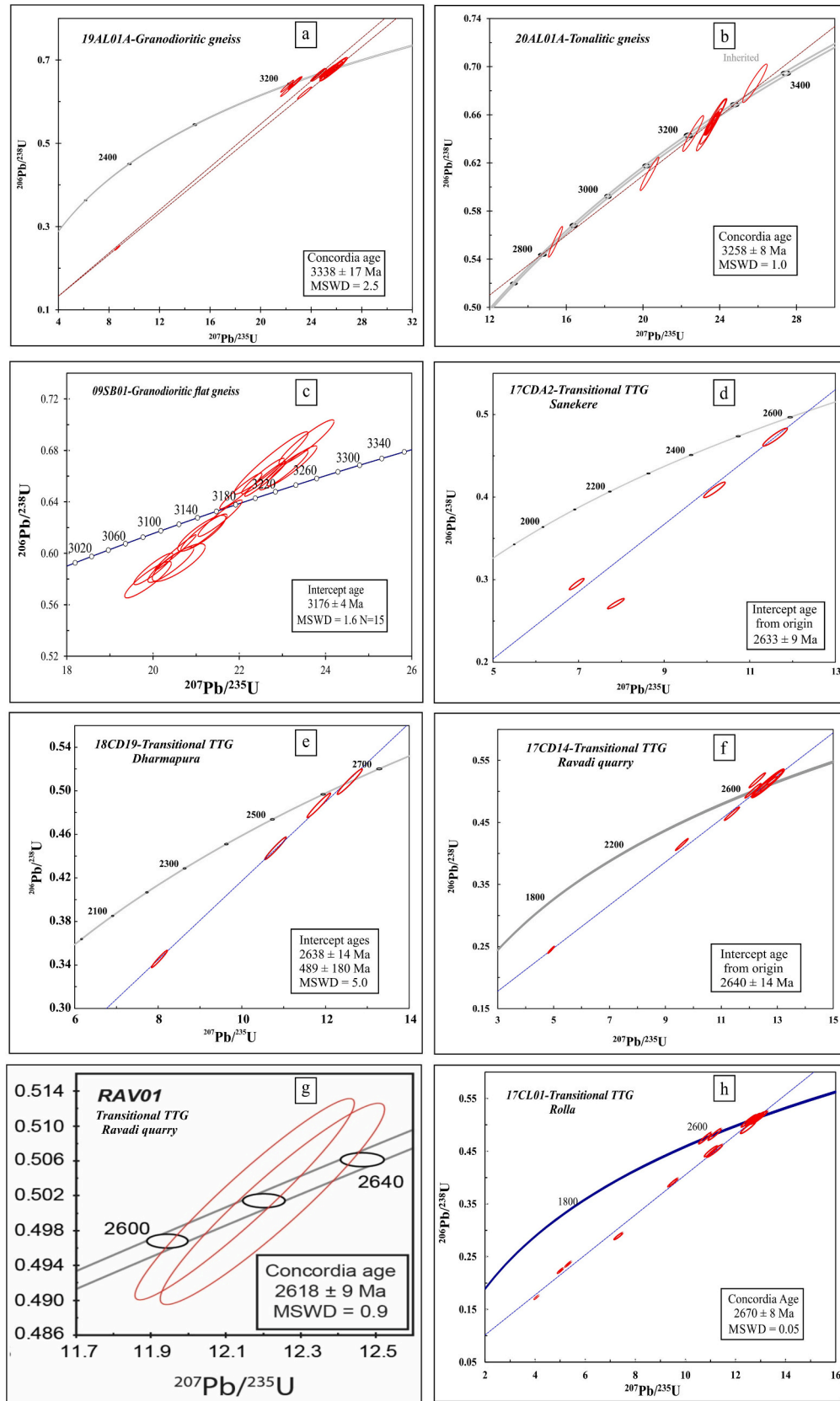


Fig. 17. Zircon U-Pb concordia diagrams of granitoids to the east of Chitradurga boundary shear zone (a) zircon U-Pb concordia diagram of granodioritic gneiss (19AL-01A) from Alappanagudda quarry (b) U-Pb concordia diagram of banded tonalitic gneiss (20AL-01A) (c) U-Pb concordia diagram of grey granodioritic flat gneiss (SB-01) (d) U-Pb concordia diagram of transitional TTG (CD-A2) (e) U-Pb concordia diagram of banded tonalitic flat gneiss (18CD-19) (f) U-Pb concordia diagram of banded grey granodioritic gneiss (17CD-14) (g) U-Pb concordia diagram of tonalitic gneiss (RAV-01) (h) U-Pb concordia diagram of banded dark grey transitional TTG (17CL-01A) (i) U-Pb concordia diagram of granodiorite (17JNK-01) (j) U-Pb concordia diagram of grey granite (18CD-21).

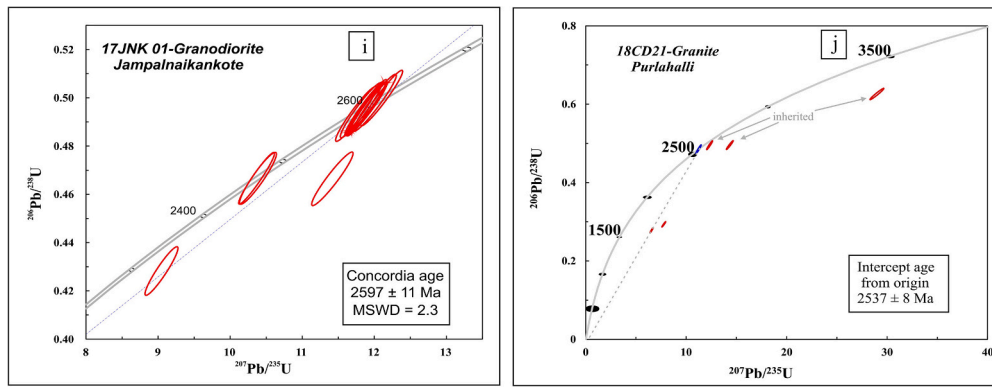


Fig. 17. (continued).

(Fig. 16). U-Pb data define a single discordia line in a concordia plot (Fig. 15d) with an upper-intercept age of 3229 ± 5 Ma (2σ MSWD = 1.14) and a lower-intercept at 743 ± 40 Ma.

6.4.5. Sample 20 WCH-06A - 3206.1 ± 8.0 Ma

Sample 20 WCH-06A medium grained fragmented dioritic dyke within tonalitic gneiss collected from Bharamanayakanadurga ($14^{\circ}11'13''\text{N}; 76^{\circ}07'43''\text{E}$). Zircon crystals from WCH-06A do not display fine-oscillatory zoning but instead present somewhat complex textures with needle-like intertwined bright and dark domains that are elongated parallel to the long axis of zircon crystals (see Fig. 16). Some bright homogeneous zones resemble metamorphic domains. U-Pb data are near-concordant and define a continuum from fully concordant data providing an age of 3206.1 ± 8.0 Ma (2σ ; MSWD = 0.26) to 5% discordant data that altogether giving an upper-intercept age of 3210.1 ± 8.7 Ma (2σ ; MSWD = 0.44) essentially projected from zero in a concordia plot (Fig. 15e). A single data point 3.2% discordant stands out because it does not fall on the continuum but rather on its left in a concordia plot. This point could originate from non-zero-age U-Pb disturbance. All crystals are characterized by very low U and Th contents of 30–98 and 10–73 ppm, respectively, and Th/U of 0.29–0.84.

6.4.6. Sample 20 WCH-07A - 3234 ± 8 Ma

Sample WCH-07A is a medium dark grey tonalitic gneiss collected from an abandoned quarry ($14^{\circ}03'27''\text{N}; 76^{\circ}13'15''\text{E}$) in Holalkere-Chitradurga road. Zircon crystals display various textures from very simple fine-oscillatory zoning to convolute zoning and dark oscillatory zoning (Fig. 16). This is also reflected in a large range of U and Th concentrations of 129–800 and 30–397 ppm, respectively. Th/U ratios returned values between 0.23 and 0.89. U-Pb data show significant discordance except for one datapoint towards which all data converges. The defined age of 3234 ± 8 Ma (2σ ; $^{207}\text{Pb}/^{206}\text{Pb}$ age; Fig. 15f) corresponds to the zircon with least U and Th content. The complex distribution of discordant data is attributed to partial resetting of the U-Pb system because all data are constrained within two discordia lines that connect the concordant datapoint to 0 and to 800 Ma.

6.5. ca. 3000 Ma potassic granite intrusions

Several discrete high potassic granite plutons intrude the basement along the western margin of Chitradurga greenstone belt.

6.5.1. Hosdurga pluton: Sample 20 WCH-05B - 2955 ± 5 Ma

Sample 20 WCH-05B is a coarse-grained light pink granite (Hosdurga pluton) collected from a canal section ($13^{\circ}53'50''\text{N}; 76^{\circ}25'57''\text{E}$) about 2 km north of Janakal village in Hosdurga – Sivaganga road. Zircon crystals commonly exhibit dark fine-oscillatory zoning from core to rim with the occurrence of several small and bright inclusions in CL images (Fig. 16). These inclusions were avoided for U-Pb dating. In line with

these dark CL images, most crystals have low to high U and Th contents of 74–1180 and 76–1305 ppm, associated with Th/U between 0.60 to 1.11. U-Pb ages are concordant to very discordant (59%) but distribute along a single discordia line in a concordia plot. The upper- and lower-intercepts gave 2955 ± 5 Ma (2σ ; MSWD = 0.20) and 455 ± 12 Ma, respectively (Fig. 15g). Our zircon U-Pb ages of basement gneiss and Hosdurga granite ages are in consistent with documented U-Pb zircon age of 3197 ± 11 Ma for the gneiss, 3010 ± 25 Ma and 3026 ± 26 Ma for Hosdurga granite by Ravindran et al. (2023).

6.5.2. Bellur-Nagamangala granite : Sample BC 01 - 2977 ± 7 Ma

Sample BC 01 is a coarse-grained whitish grey monzogranite collected from a quarry ($12^{\circ}58'33''\text{N}; 76^{\circ}46'52''\text{E}$) near Bellur cross (see Fig. 1b) in Bangalore – Mangalore National Highway. Zircon crystals are euhedral, stubby to slightly elongated with oscillatory zoning (Fig. 16). In some grains patches of bright domains partly distorting oscillatory zoning can be observed. Most zircons contain low to moderate U and Th contents 20–114 ppm, 21–185 ppm, respectively with Th/U ratios (0.45–1.72). The determined least discordant U-Pb data indicates an upper-intercept age of 2977 ± 7.0 Ma (2σ ; MSWD = 0.8) which we interpret as that of granite crystallization (Fig. 15h).

6.6. Eastern corridor - TTG-type migmatitic gneisses (3338–3176 Ma)

In this study, we have analysed zircons from older TTG-type migmatitic gneisses and younger banded transitional TTGs to the east of Chitradurga boundary shear zone.

6.6.1. Sample 19 AL-01A – Granodioritic gneiss: 3338 ± 17 Ma

Sample 19 AL-01A is a coarse-grained weak grey granodioritic gneiss collected from Alappanagudda quarry in Kunigal-Yediyur road ($13^{\circ}00'19''\text{N}; 76^{\circ}57'35''\text{E}$).

Zircon crystals from 19 AL 01A mostly exhibited fine-oscillatory zoning but several crystals also showed convolute zoning and bright homogeneous domains crosscutting parts of oscillatory-zoned domains, in a way similar to zircons from 20-AL-01-A (Fig. 16). All crystals are characterized by low U and Th contents of 16–151 and 7–92 ppm, respectively. Th/U ratios are quite variable because they range from 0.09 to 0.89 (Supplementary table 2). U-Pb data have somewhat complex distribution in Concordia plot but define a maximum age of 3338 ± 17 Ma. They are distributed along the Concordia curve, much like in the case of 20-AL-01-A, though with some datapoints being more discordant. We interpret the data as reflecting magmatic crystallization at 3338 ± 17 Ma (2σ ; MSWD = 2.5; Fig. 17a), ancient metamorphic partial resetting of the U-Pb isotope system around 3000–2800 Ma, and later alteration resulting in nearly zero-age discordance. This age is consistent with published U-Pb zircon ages along the eastern margin of Javagondanahalli further north.

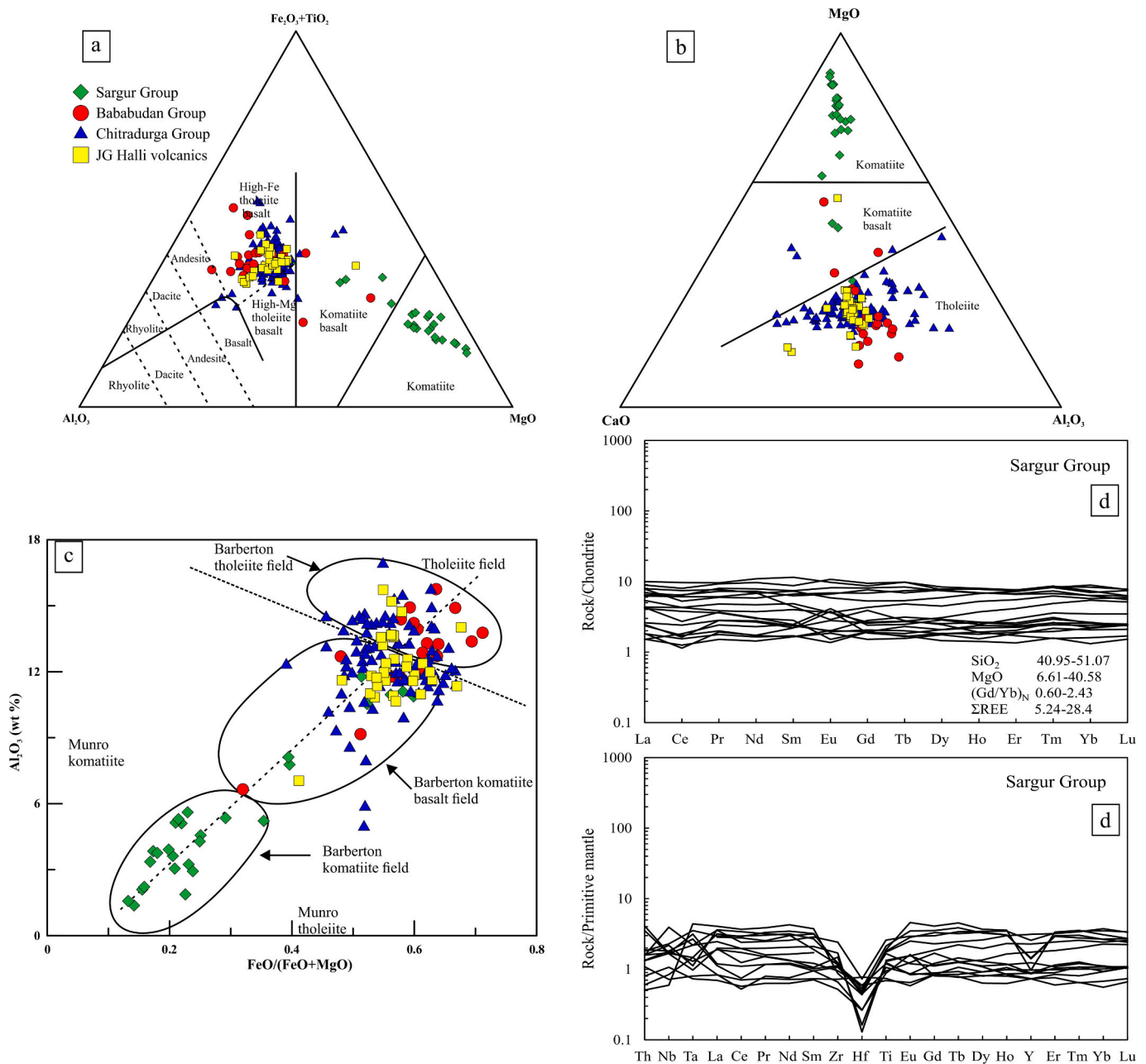


Fig. 18. (a) $\text{Al}_2\text{O}_3\text{-Fe}_2\text{O}_3+\text{TiO}_2\text{-MgO}$ (AFM) ternary plot (Jensen, 1976 modified by Viljoen et al., 1982) (b) $\text{CaO-MgO-Al}_2\text{O}_3$ ternary plot (Viljoen et al., 1982). (c) $\text{FeO}/(\text{FeO}+\text{MgO})$ versus Al_2O_3 (Arndt, 2003) binary diagram showing Barberton komatiite and komatiite basalt to tholeiite field for the Sargur Group and Bababudan, Chitradurga, J.G. Halli respectively (d) Chondrite normalized REE plot and primitive mantle normalized (Sun and McDonough, 1989) multi-element spider diagrams of Sargur Group volcanics.

6.6.2. Sample 20 AL-01A – Tonalitic gneiss: 3258 ± 8 Ma

Sample 20 AL-01A collected from Alappanagudda quarry ($13^\circ 00' 19''\text{N}; 76^\circ 57' 35''\text{E}$) is a coarse grained dark grey banded dioritic gneiss found as large disrupted boudins along the shear zone that traverse the weakly foliated granodioritic gneiss.

Sample 20 AL-01A has zircon crystals that are characterized by fine-oscillatory zoning from core to rim but in numerous cases bright homogeneous zones crosscut the fine-oscillatory zoned domains (Fig. 16). This is typical of metamorphic recrystallization (Pidgeon, 1992; Corfu and Hegde, 2020). Zircon U-Pb data for this sample is distributed along the concordia in between 3350 and 2800 Ma. This distribution is interpreted as reflecting igneous crystallization at 3258 ± 8 Ma (2σ ; MSWD = 1; Fig. 17b) and U-Pb partial resetting around 2800 Ma, which was accompanied by metamorphic recrystallization. All of analyzed

zircons displayed low U and Th concentrations of 21-120 and 11-88 ppm, respectively, with relatively reproducible Th/U of 0.50-0.67 except for two analyses that gave 0.29 and 1.04 (Supplementary table 2.).

6.6.3. Sample SB-01 – Granodioritic flat gneiss: 3176 ± 4 Ma

Sample SB-01 is a coarse- to medium-grained light grey granodioritic flat gneiss collected from the road cut 1 km NNW of Seebi temple in the Bangalore -Pune National highway ($12^\circ 38' 58''\text{N}; 76^\circ 59' 33''\text{E}$). Zircon crystals are euhedral to subhedral with fine oscillatory zoning structures from core to rim (Fig. 14). Most zircon grains exhibit a large range of U and Th contents (low to high) of 45-909 ppm and 11-468 ppm respectively. U-Pb data of 15 grains near concordant age define 3176 ± 4 Ma with (2σ ; MSWD = 1.6; Fig. 17c). We interpret this age as a

crystallization stage of their magmatic protoliths. This age coincides with ages of magmatic basement further south in Bangalore region (Jayananda et al., 2000).

6.7. Transitional TTGs (2669-2618 Ma)

6.7.1. Sample 17CD-A2 - Transitional TTGs: 2633 ± 9 Ma

Sample 17CD-A2 is coarse grained banded transitional TTG collected close to Sanekere ($14^{\circ}10'58''\text{N}; 76^{\circ}39'53''\text{E}$) in Hiriya - Challakere road. The rock contains quartz, plagioclase (An_{18-24}), minor microcline, hornblende, biotite with accessory zircon, apatite, titanite and opaque phases. Most 17 CD-A2 zircons are characterized by simple and fine oscillatory zoning from core to rim with some of them have a very bright rim in CL images (Fig. 16), although this sample yielded few zircons of good enough quality to attempt U-Pb dating. These crystals have moderate to high U concentrations of 262-622, low Th concentrations of 53-91 ppm, and Th/U of 0.15-0.34 (Supplementary table 2). All U-Pb data are discordant but they somewhat converge towards the least discordant datapoint which has a degree of discordance of 6.2% and an age of 2633 ± 9 Ma (2σ ; MSWD = 4.3) when projected from the origin (Fig. 17d).

6.7.2. Sample 18 CD 19 - Transitional TTG: 2638 ± 14 Ma

Sample CD 19 is a dark grey banded tonalitic flat gneiss from Dharmapura ($14^{\circ}05'34''\text{N}; 76^{\circ}50'40''\text{E}$). This sample comprises quartz, abundant plagioclase, minor microcline with hornblende, biotite with common accessory zircon, apatite, titanite and opaques in the mineral assemblage. Sample 18 CD 19 yielded few zircons among which only crystals exhibited igneous textures characterized by fine oscillatory zoning from core to rim (Fig. 16). Some darker zones were nevertheless observed in some crystals but they were concentrically distributed which likely is a result of lattice damaging from U and Th decay. Low to moderate U and Th concentrations were observed in these zircons (144-407 and 46-255 ppm, respectively). Determined Th/U ratios range from 0.19 to 0.63 and do not correlate with ages, nor degrees of discordance, which range from -0.1 to 28.6% (Supplementary table 2). Overall data align on a discordia line that has an intercept age of 2638 ± 14 Ma (2σ ; MSWD = 5) with the most concordant zircon being 2647 ± 17 Ma (Fig. 17e).

6.7.3. Sample 17 CD 14 - Transitional TTG: 2640 ± 14 Ma

Sample 17CD 14 is a coarse-grained banded grey granodioritic gneiss

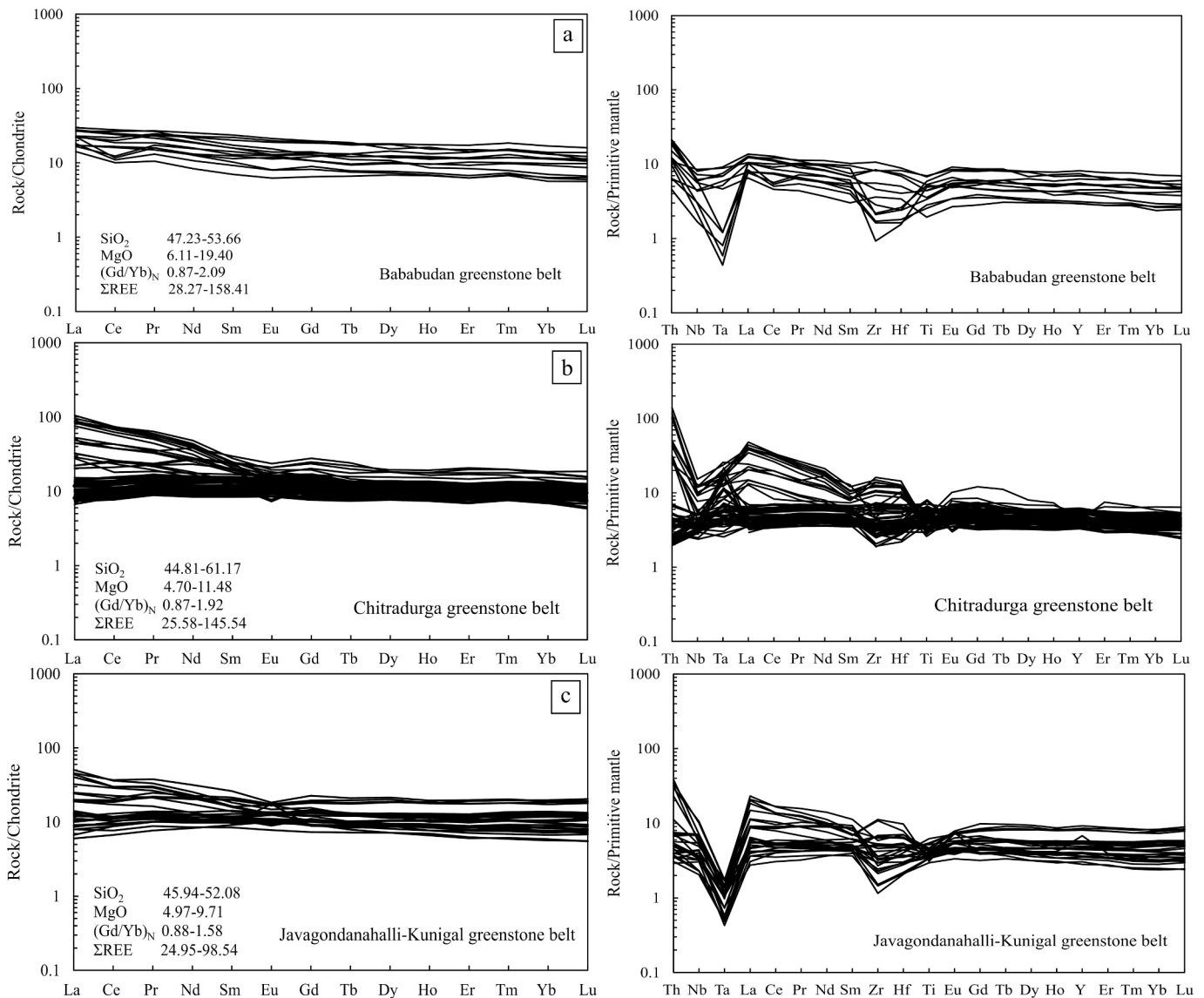


Fig. 19. (a) Chondrite normalized REE plot and Primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of Bababudan greenstone belt (b) Chondrite normalized REE plot and Primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of Chitradurga greenstone belt (c) Chondrite normalized REE plot and Primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of Javagondanahalli-Kunigal greenstone belt.

from Ravadi quarry (13°51'30"N;77°00'53"E). The rock contains quartz, plagioclase (An₁₆₋₂₂), subordinate microcline, hornblende and biotite with accessory zircon, apatite, titanite and opaques phases. Zircons occur as elongated prismatic grains which show oscillatory zoning with evidence for recrystallization along the margin locally observed in CL images (Fig. 16).

The zircon crystals recovered from 17 CD 14 all exhibit fine oscillatory zoning from core to rim, though with some bright domains in CL that are irregular and remind metamorphic recrystallization fronts (Fig. 17f). Concentrations of U and Th vary from low (124 and 60 ppm, respectively) to high (715 and 462 ppm, respectively) which correspond to Th/U of 0.37 to 0.74 (Supplementary table 2). Most U-Pb data are concordant to slightly discordant are consistent with a Concordia age of 2640 ± 14 Ma (2 σ ; MSWD= 1.4; Fig. 17f) which corresponds to magmatic crystallization of zircon.

6.7.4. Sample RAV 01 - Transitional TTG: 2618 ± 9 Ma

Sample RAV 01 is a dark grey tonalitic gneiss from Ravadi quarries (13°51'30"N; 77°00'53"E). The rock is medium-grained, show strong fabrics and contains quartz, plagioclase (An₂₂₋₂₈), minor microcline,

hornblende, biotite, remnants of clinopyroxene with accessory zircon, apatite and opaque phases.

This sample has only yielded one zircon with sufficiently low radiation damage features to be confidently analyzed for U-Pb geochronology. The crystal exhibits a bright core with faint zoning with homogeneous dark grey outer parts with uneven contact with the core likely representing metamorphic recrystallization (Fig. 16). Analyses were done in the core which preserved an igneous aspect. The RAV01 crystal was dated twice and had low U and Th contents of 74-267 ppm and 85-171 ppm, respectively, which translates into Th/U of 0.64-1.15 (Supplementary table 2). Determined U-Pb age gave an average of 2618 ± 9 Ma (2 σ ; n = 2; MSWD = 0.9; Fig. 17g) out of the two concordant analyses.

6.7.5. Sample 17 CL 01A - Transitional TTG: 2669 ± 7.7 Ma

Sample 17 CL 01A is a weakly banded dark grey transitional TTG collected from the western margin of the Closepet batholith (Jayananda et al., 2020) about 4 km north of Rolla village (13° 52'32"N; 77°05'35"E). The sample contains quartz, plagioclase (An₁₇₋₂₅), microcline, biotite, hornblende with accessory phases of zircon, apatite,

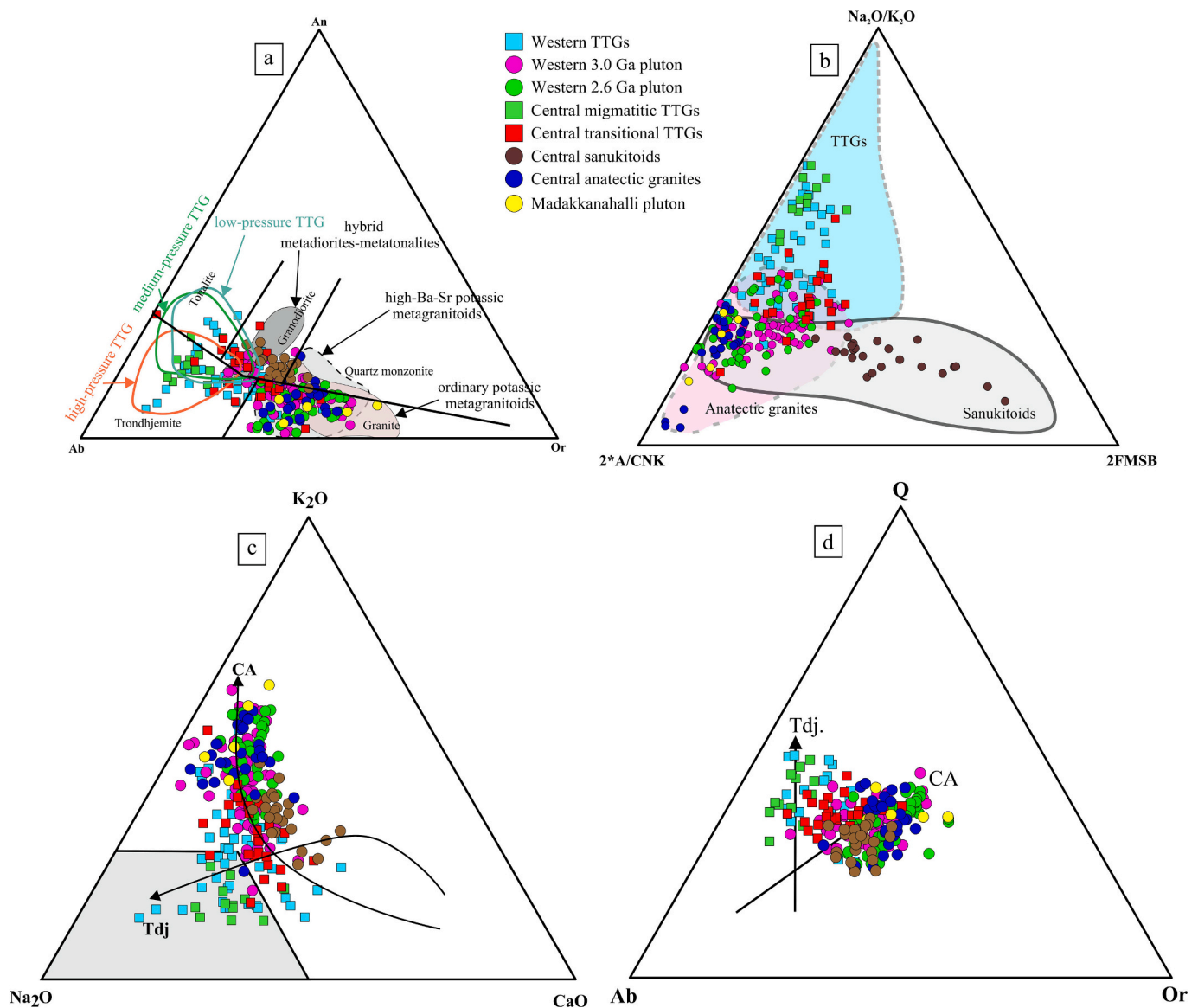


Fig. 20. (a) Normative Ab-An-Or triangular diagram (after O'Connors, 1965) with fields from Barker (1979) (b) 2*A/CNK-Na₂O/K₂O-2FMSB triangular plot (Laurent et al., 2014) (c) Na-K-Ca triangular plot (Barker and Arth, 1976) (d) normative Q-Ab-Or (Barker and Arth, 1976).

titanite and opaques. Zircon grains are elongated, euhedral and CL images reveal oscillatory zoning indicating their magmatic origin (Fig. 16). Some bright areas together with darker domains appears to be resulted by radiation damage induced metamictization. Some brighter domains reflect effect of metamorphic recrystallization probably linked to emplacement of the Closepet batholith.

Zircon crystals from 17 CL 01A have been studied in Jayananda et al. (2020b). They show simple fine-oscillatory zoning from core to rim and returned concordant U-Pb data that correspond to a concordia age of 2669.9 ± 7.7 Ma (2σ ; MSWD = 1.5; Jayananda et al., 2020). This age is interpreted as that of zircon population crystallization. Concentrations of U and Th are relatively low (154–228 and 49–85 ppm, respectively) and Th/U show limited variability (i.e., 0.28–0.54) within the typical magmatic range (i.e., ~ 0.2 –0.8).

6.8. Calc-alkaline granitoids (2600–2500 Ma)

6.8.1. Sample 17 JNK 01- Granodiorite : 2597 ± 11 Ma

Sample 17 JNK 01 is a light grey granodiorite collected from an abandoned quarry in the southern outskirts of Jampalnakankote ($14^{\circ}10'28''$ N; $76^{\circ}32'58''$ E).

Zircon crystals from 17 JNK 01A sample commonly show somewhat complex internal structures made of fine oscillatory zoning whereas other domains show convolute zoning and/or dark and patchy aspects suggesting high levels of radiation damage accumulation (see Fig. 16). The most concordant U-Pb data indicates a concordia age of 2597 ± 11 Ma (2σ ; MSWD = 2.3; Fig. 17i). Concentrations of U and Th vary from low (93 and 11 ppm, respectively) to high (1191 and 1429 ppm, respectively) which correspond to Th/U of 0.02 to 1.2 (Supplementary table 2).

6.8.2. Sample CD 21- Granite: 2537 ± 8 Ma

Sample CD 21 is light medium to fine grained grey granite cutting

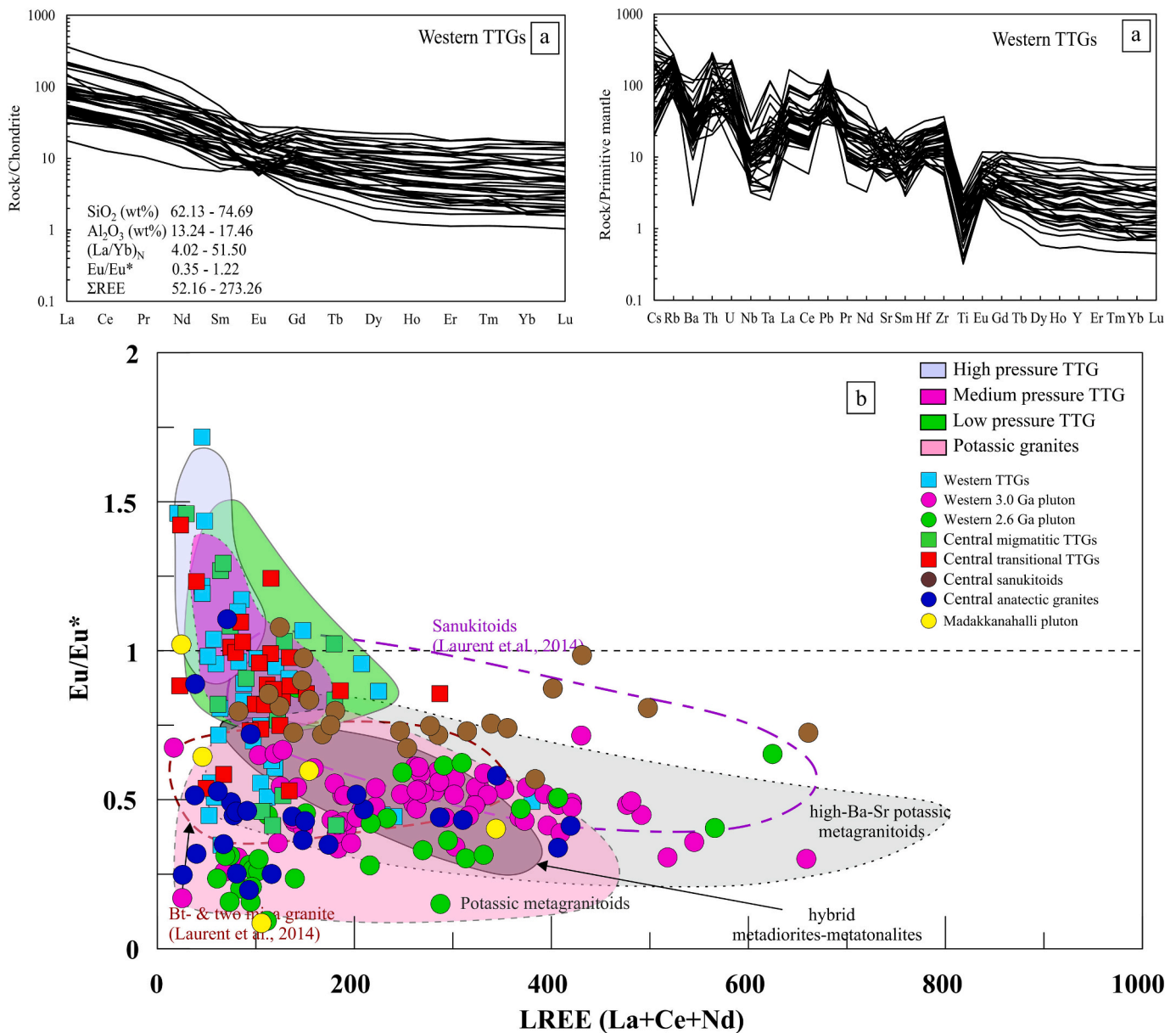


Fig. 21. (a) Chondrite normalized REE plot and primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of western TTGs (b) Eu/Eu^* versus LREE plot (after Simon et al., 2021).

across flat gneisses collected at Purlahalli (14°10'13"N;76°50'40"E).

Sample 18 CD21 contains zircons that for the most part exhibit fine oscillatory zoning from core to rim in CL images with, however, some darker domains, which likely result from radiation damage accumulation, and small convolute brighter zones that crosscut zoning, suggesting metamorphic recrystallization (see Fig. 16). Some crystals have inherited cores based on CL images because they have a different shade of grey and internal zoning that is inconsistent with that of the rest of the crystal with which a zoning crosscut is evident (see Fig. 16). This is shown in Wetherill concordia plot which indicates that the age of the igneous zircon population within 18 CD 21 is 2537 ± 8 Ma (2σ ; MSWD = 0.56; Fig. 17j). The oldest inherited zircon has a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3644 ± 16 Ma but the low $^{206}\text{Pb}/^{204}\text{Pb}$ of 78 together with its high U and Th contents of 1895 and 1357 ppm, respectively, would suggest a common Pb influence which could significantly overestimate the age (Supplementary table 2).

7. Chemical characteristics of greenstone – granitoid assemblages

Major and trace element data of greenstone volcanics (from different

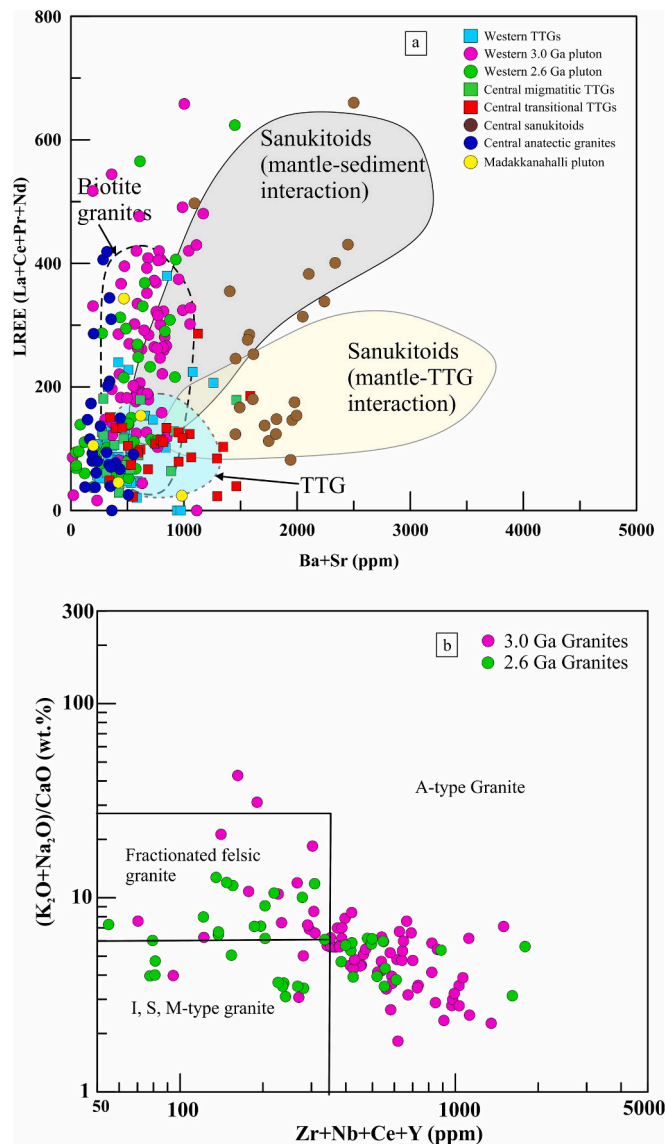


Fig. 22. (a) LREE versus Ba+Sr binary plot (Laurent et al., 2014) (b) $(\text{K}_2\text{O}+\text{Na}_2\text{O})/\text{CaO}$ versus Zr+Nb+Ce+Y binary plot (Whalen et al., 1987).

tectono-stratigraphic units), diverse granitoids including TTGs, potassic granites, transitional TTGs, sanukitoids and anatectic granites from published record combined with new results of the present study including greenstone volcanics, granitoids are presented as Appendices (supplementary tables 3 and 4).

7.1. Greenstone volcanics: major and trace elements

Greenstone volcanics from different tectono-stratigraphic units display diverse elemental characteristics and ratios (Jayananda et al., 2008, 2016; Tushipokla and Jayananda, 2013; Bhattacharyya et al., 1988; Duraiswami et al., 2013; Sreehari et al., 2021; Rose et al., 2022; Manikyamba et al., 2022).

7.1.1. Sargur Group greenstone volcanics (ca.3400-3200 Ma)

Published elemental data together with our new data on the Sargur Group volcanic rocks from the western boundary of the Chitradurga greenstone belt indicate mainly ultramafic composition with few samples extending into mafic domain (Jayananda et al., 2016; present study). The compositions of ultramafic volcanic rocks reveal komatiite to komatiite basalt lineage with low SiO_2 (40-47 wt.%), Al_2O_3 (1.41-6.73 wt.%), MgO (15.20-40.58 wt.%) whilst mafic rocks show high-Mg basalt composition with SiO_2 (47.30-51.07 wt.%), MgO (6.61-12.3 wt.%), high Al_2O_3 (7.62-11.78 wt.%). On the $\text{Al}_2\text{O}_3\text{-Fe}_2\text{O}_3\text{-TiO}_2\text{-MgO}$ (Fig. 18a) triangular plot (Jensen, 1976 modified by Viljoen et al., 1982) and $\text{CaO-MgO-Al}_2\text{O}_3$ (Viljoen et al., 1982) the volcanic rocks mainly plot in komatiite with a few samples extend to komatiite basalt field (Fig. 18b). On the $\text{FeO}/\text{FeO}+\text{MgO}$ versus Al_2O_3 plot (Arndt, 2003) the Sargur Group volcanic rocks plot on Barberton komatiite to Barberton komatiite basalt field (Fig. 18c). The komatiites mainly display Al-depleted character with low $\text{Al}_2\text{O}_3/\text{TiO}_2$ (6.7 – 17.1 except few samples showing anomalously high values up to 32) but conversely high $\text{CaO}/\text{Al}_2\text{O}_3$ (2.5-1.03 except few samples displaying anomalous low values of 0.32 to 0.21) ratios whilst komatiitic basalts and high-Mg basalts show transitional from Al-depleted to Al-undepleted character ($\text{Al}_2\text{O}_3/\text{TiO}_2 = 17.4$ to 19.0; $\text{CaO}/\text{Al}_2\text{O}_3 = 1.05 - 0.82$). To summarize, Sargur Group volcanic rocks are dominantly komatiite composition belong to Al-depleted Barberton komatiite type with few komatiitic basalt to high-Mg basalts belong to Al-undepleted Munro-type (Arndt et al., 2008). On Harker's binary diagrams, major elements of komatiites and komatiitic basalts show strong to moderate fractionation trends wherein SiO_2 , Al_2O_3 , TiO_2 , Fe_2O_3 , CaO , show negative correlation with MgO whilst K_2O and P_2O_5 show scattering (figure not given).

Among trace elements, komatiite and komatiitic basalts are characterized by high compatible elements contents (Ni, Cr), variable but generally low content of incompatible elements Rb, Ba, Sr, U, Th, Zr, Hf, REE (supplementary table 3). On Harker's binary diagrams trace elements exhibit moderate to weak trends against MgO . Among the trace elements Ni show moderate positive correlation indicate olivine fractionation whilst Zr, Y show weak positive trend and Cr, V, Sr, Ba show scattering (figure not given).

The Sargur Group komatiites and komatiite basalts contain very low to moderate total REE (5.24-28.4 ppm). Chondrite (Masuda et al., 1973) normalized REE display slightly fractionated to flat REE patterns [$(\text{Gd}/\text{Yb})_N = 0.60\text{-}2.43$] with few samples exhibiting either negative or positive Ce and Eu anomalies (Fig. 18d). On the primitive mantle-normalized multi-element diagrams the komatiites and komatiite basalts show slight positive or no significant Nb anomalies whilst strong negative Hf anomalies, with negative or without any Y anomalies (Fig. 18d).

7.1.2. Bababudan Group (ca.3000-2720 Ma)

The Bababudan Group volcanics comprise voluminous mafic rocks with minor ultramafic rocks and intermediate rocks. Major element data reveal significant variation in SiO_2 (47.23-53.66 wt.%), MgO (6.11-19.40 wt.%). On the $\text{Al}_2\text{O}_3\text{-Fe}_2\text{O}_3\text{-TiO}_2\text{-MgO}$ (see Fig. 18a) triangular

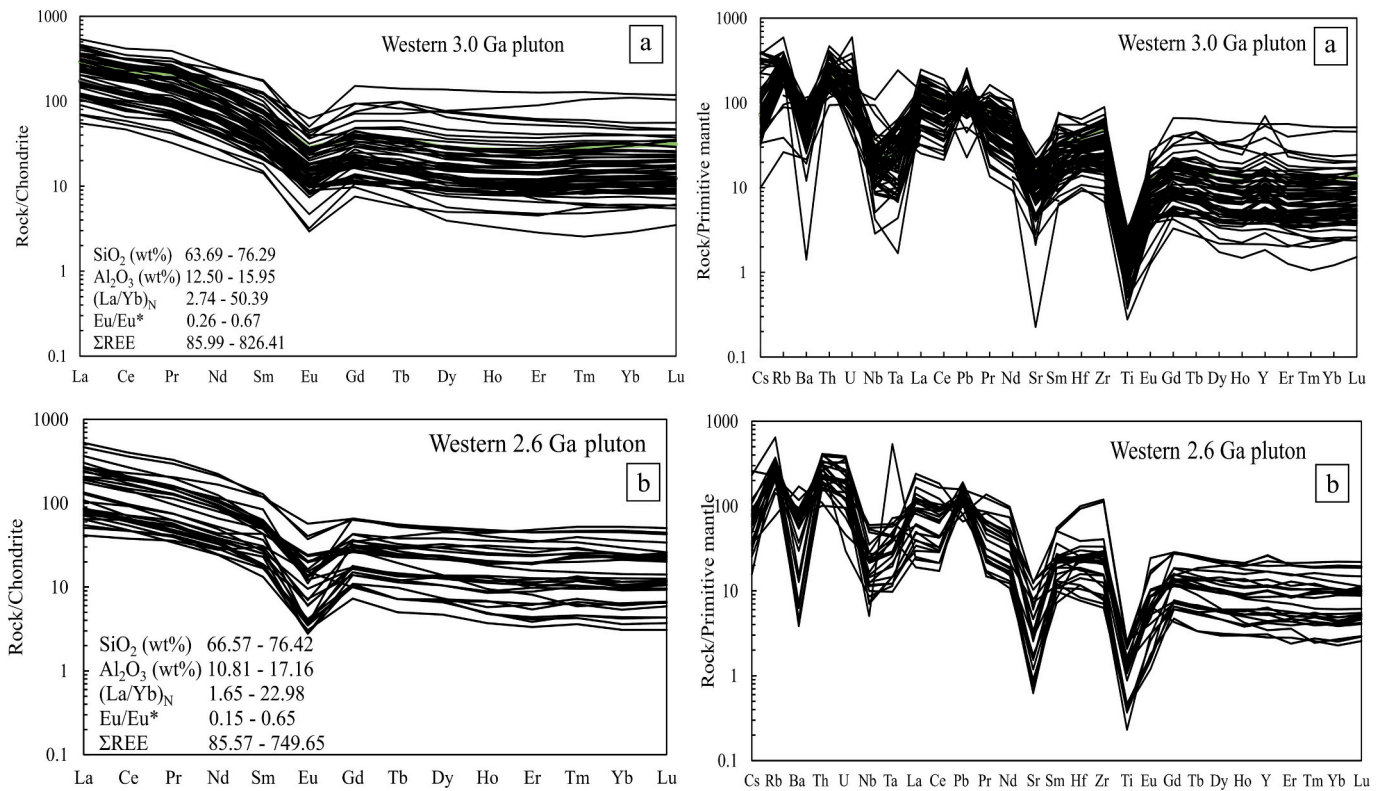


Fig. 23. (a) Chondrite normalized REE plot and primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of 3.0 Ga granites (b) Chondrite normalized REE plot and primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of 2.6 Ga granites.

plot (Jensen, 1976 modified by Viljoen et al., 1982) and CaO-MgO-Al₂O₃ (Viljoen et al., 1982) the volcanic rocks mainly plot in Fe-tholeiite field with a few samples extend to basaltic andesite and komatiite basalt field (see Fig. 18b) whilst on the CaO-MgO-Al₂O₃ (Viljoen et al., 1982) the volcanic rocks plot in the tholeiitic field with few samples fall on komatiite basalt field. On the Harker's binary plots major elements to moderate to weak trends against MgO wherein SiO₂, Al₂O₃, TiO₂, Fe₂O₃, Na₂O, P₂O₅ exhibit moderate negative correlation whilst K₂O show large scattering (figure not given). Among trace elements Ni and Cr show positive correlation whilst Zr, Y and Sr display negative correlation and Ba exhibit large spread. The Bababudan Group volcanics show large variation in total REE contents (28.27-158.41 ppm) and chondrite normalized REE display flat to slightly fractionated patterns [(Gd/Yb)_N = 0.87-2.09] with a few samples show negative Ce and Eu anomalies which probably linked to hydrothermal activity (Fig. 19a). On the primitive mantle (Sun and McDonough, 1989) multi-element diagrams these volcanics show large variation wherein about 50% of the samples exhibit strong negative Nb, Ta, Zr, Hf, anomalies whilst remaining samples display moderate to weak negative anomalies. Two anomalous samples exhibit strong positive Ta, Zr, Hf anomalies (see Fig. 19a).

7.1.3. Chitradurga group (ca. 2746-2667 Ma)

The Chitradurga greenstone volcanics contains dominant mafic with subordinate intermediate volcanics to felsic pyroclastic flows. They show wide range of major element compositions with SiO₂ (44.81 to 61.17 wt.%), MgO (4.70 -11.48 wt.%) and Al₂O₃ (7.93-16.87 wt.%). Al₂O₃ (6.67 to 16.39). On Al₂O₃-Fe₂O₃+TiO₂ -MgO (see Fig. 18a) triangular plot (Jensen, 1976 modified by Viljoen et al., 1982) the volcanic rocks mainly plot in high-Fe-tholeiite field with a few samples extend to high-Mg tholeiite and komatiite basalt field (see Fig. 18a) whilst on the CaO-MgO-Al₂O₃ (Viljoen et al., 1982) the volcanic rocks plot on tholeiite field with three samples fall on komatiite basalt field (see Fig. 18b). Major element oxides on Harker's binary diagrams do not

define differentiation trends instead form cluster or large spread (figure not given). Published data together with our new data on the Chitradurga volcanics reveal low to high total REE (25.58- 145.54 ppm). Chondrite normalized REE show flat to moderately fractionated REE patterns with (Gd/Yb)_N ratios ranging 0.87-1.92 with few samples show negative Eu anomalies (Fig. 19b). On the primitive mantle normalized multi-element plot 2/3 of studied samples show negative Nb-Ta-Ti-Sr anomalies with remaining samples displaying positive Ta-Ti-Sr (see Fig. 19b). On the other hand, majority Chitradurga samples exhibit positive Th, U, La, Ce, Zr, Hf anomalies.

7.1.4. Javagondanahalli volcanics (ca. 2740 Ma)

The Javagondanahalli together with Kunigal greenstone volcanics are dominated by amphibolites with minor tuffs. They display significant variation in SiO₂ (45.94-52.08 wt.%), MgO (4.97-9.71 wt.%), Al₂O₃ (7.06-15.71 wt.%) and Fe₂O₃ (10.56-13.81 wt.%).

On the Al₂O₃-Fe₂O₃+TiO₂ -MgO (see Fig. 18a) triangular plot (Jensen, 1976 modified by Viljoen et al., 1982) and CaO-MgO-Al₂O₃ (Viljoen et al., 1982) the amphibolite mainly plot in Fe-tholeiite field with one sample extending into komatiite basalt field (see Fig. 18b) whilst on the CaO-MgO-Al₂O₃ (Viljoen et al., 1982) they plot on tholeiite except one sample plot on komatiite basalt field. On Harker's binary plots major element oxides form cluster and do not define a differentiation trends. The Javagondanahalli- Kunigal amphibolite samples contain moderate to high contents of REE (ΣREE = 24.95-98.54 ppm). They display flat to slightly fractionated REE patterns with (Gd/Yb)_N ratios ranging 0.88 to 1.58 and slightly negative to positive Eu anomalies (Fig. 19c). On the primitive mantle normalized (Sun and McDonough, 1989) majority of these rocks exhibit variable negative Nb, Ta, Zr, Hf anomalies (see Fig. 19c). On the other hand, Javagondanahalli samples show positive Sr anomalies whilst majority of Kunigal samples show negative Sr anomalies except three samples exhibit slight positive anomalies.

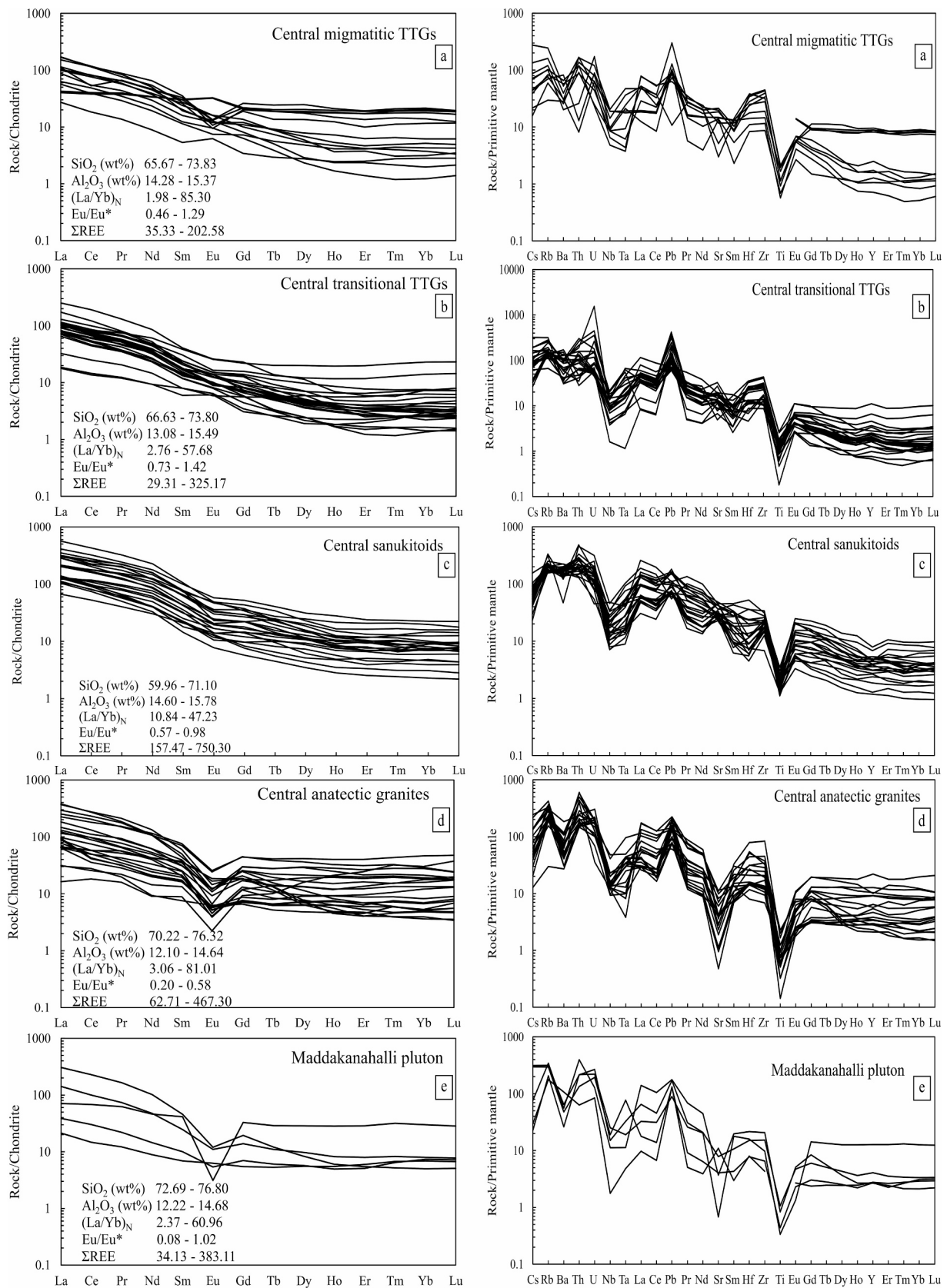


Fig. 24. (a) Chondrite normalized REE plot and primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of central migmatitic TTGs (b) Chondrite normalized REE plot and Primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of transitional TTGs (c) Chondrite normalized REE plot and primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of sanukitoids (d) Chondrite normalized REE plot and primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of anatectic granites (e) Chondrite normalized REE plot and primitive mantle normalized (Sun and McDonough, 1989) multi-element plot Maddakanahalli granite.

8. Archean granitoids across the boundary shear zone: major and trace element characteristics

The granitoids of the studied corridor across the Chitradurga boundary shear zone show wide spectrum of major and trace element compositions ranging from TTGs to transitional TTGs, high-potassic granites, sanukitoids and anatectic granite (Jayananda et al., 2018). Here we present our new major and trace element data of granitoids (Supplementary table 4) together with published record (Jayananda et al., 2006, 2018, 2019, 2020; Mohan et al., 2023; Ravindran et al., 2021, 2024; Sebastian et al., 2021, 2023; Ao et al., 2021).

8.1. Western corridor - TTG-type granitoids

The TTG-type granitoids show wide ranges of major element compositions [$\text{SiO}_2 = 62.13$ to 74.69 wt.%; $\text{Al}_2\text{O}_3 = 13.24$ – 17.46 wt.%; $\text{Na}_2\text{O} = 3.70$ – 5.61 wt.%; $\text{K}_2\text{O} = 1.32$ – 2.91 wt.%; $\text{MgO} = 0.18$ – 2.78]. These

gneisses are metaluminous to peraluminous ($0.7 \leq \text{A/CNK} \leq 1.3$; Figure is not given). They display K_2O -poor and Na_2O -rich character with high $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios (1.1 to 4.4). On Ab-An-Or triangular plot (after O'Connor, 1965) these granitoids plot on mainly on trondhjemite and extending into tonalite and granodiorite to granite fields (Fig. 20a) whilst on the $2^*\text{A}/\text{CNK}-\text{Na}_2\text{O}/\text{K}_2\text{O}-2\text{FMSB}$ triangular plot they plot (Laurent et al., 2014) they plot on the TTG field (Fig. 20b). Further, they define trondhjemite differentiation trend on the K-Na-Ca plot (Fig. 20c; Barker and Arth, 1976) and on normative Q-Ab-Or plot (Barker and Arth, 1976; Fig. 20d) which is typical of Archean TTGs (Martin and Moyen, 2002). Their Al_2O_3 contents reveal both high-Al ($\text{Al}_2\text{O}_3 > 14.0$ wt.%) and low-Al ($\text{Al}_2\text{O}_3 < 14.0$ wt.%) type TTGs. On Harker's binary diagrams major elements show moderate to strong differentiation trends wherein Al_2O_3 , TiO_2 , Fe_2O_3 , MgO , CaO , P_2O_5 exhibit negative correlation whilst Na_2O and K_2O show large spread (Figure is not given).

The TTGs are characterized by low to high content of incompatible trace elements (Rb, Ba, Sr, Zr, Hf, Th, U and REE) but conversely low

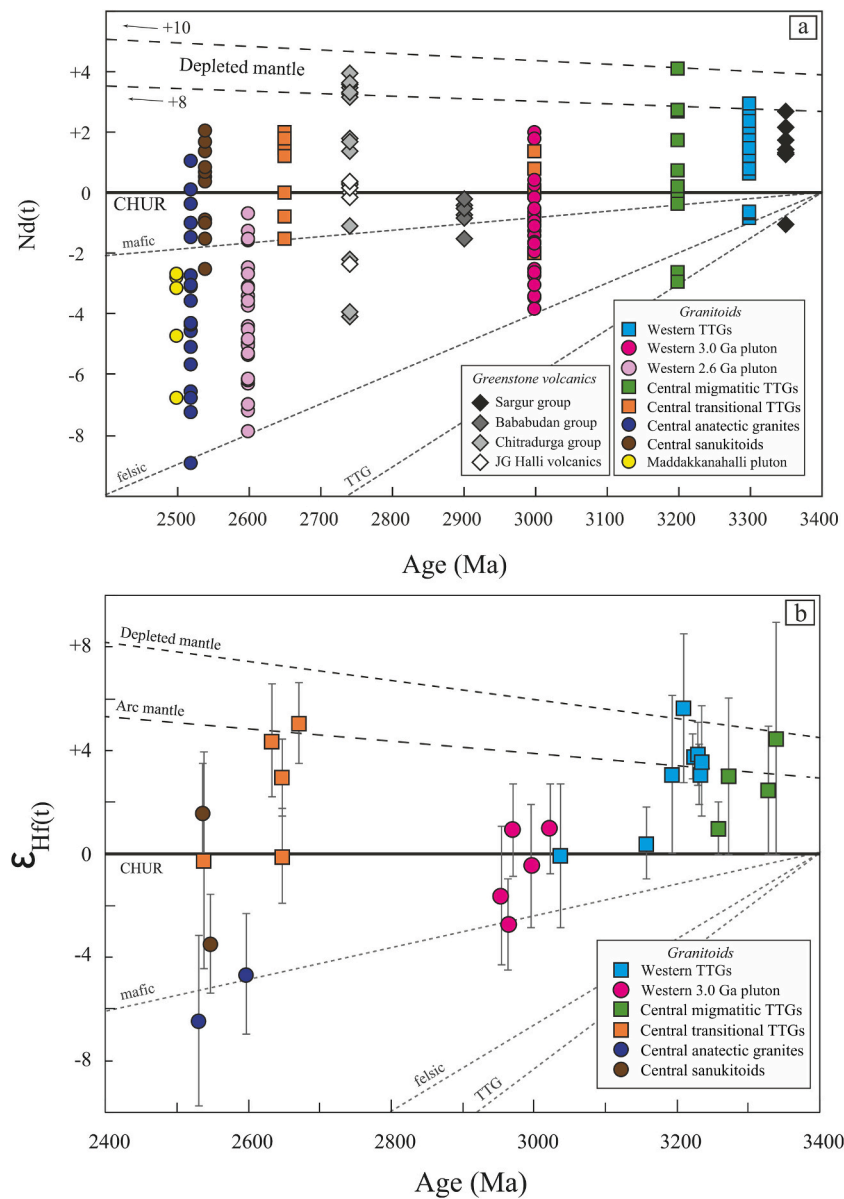


Fig. 25. (a) $\epsilon_{\text{Nd}}(t)$ versus age evolution diagram of greenstones, TTGs from western and central blocks showing involvement of depleted mantle with traces of ancient crustal contamination (b) $\epsilon_{\text{Hf}}(t)$ versus time evolution diagram of western TTGs, 3.0 Ga and 2.6 Ga granites and central migmatitic, transitional TTGs, sanukitoid and anatectic granite showing involvement of depleted mantle with variable amount of pre-existing crust. Mafic, felsic, and TTG evolutionary trajectories have been calculated using $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{176}\text{Lu}/^{177}\text{Hf}$ of, respectively, 0.18 and 0.025, 0.12 and 0.010, and 0.08 and 0.004.

abundances of compatible trace elements (Ni, Cr) except few samples show higher Cr (>100 Ma and Ni >10). They exhibit variable total REE (52.16 - 273.26 ppm) concentrations which display poor to highly fractionated REE patterns $[(La/Yb)_N = 4.02 - 51.50]$ with minor negative or without any significant Eu anomalies except few samples show positive Eu anomalies ($Eu/Eu^* = 0.35 - 1.22$) (Fig. 21a). These granitoids plot on field defined for TTGs on Eu/Eu^* versus LREE plot (Fig. 21b, after Simon et al., 2021) and LREE versus Ba+Sr plot (Fig. 22a; Laurent et al., 2014). These TTGs on the primitive mantle (Sun and McDonough, 1989) normalized multi-element diagrams they show prominent negative Ba, Nb-Ta, Ti, Sm and Sr except few samples exhibit positive Sr anomalies.

8.1.1. ca.3000-2950 Ma potassic granites

The ca. 3000 Ma old Bellur-Nagamangala, Bukkapatna, Hosdurga plutons exhibit euhedral quartz and primary epidote suggesting their rapid emplacement into the basement. They exhibit wide range of major element compositions [$SiO_2 = 63.69$ to 76.29 wt.%; $Al_2O_3 = 12.50$ - 15.95 wt.%; $Na_2O = 4.56$ - 5.61 wt.%; $K_2O = 3.96$ - 5.68 wt.%; $MgO = 0.14$ - 1.44 wt.%]. Majority of the samples are peraluminous ($0.9 \leq A/CNK \leq 1.3$) with few samples extending to metaluminous field (Figure is not given). On the Ab-An-Or plot (see Fig. 20a; O'Connor, 1965) they plot on granite field with few samples extending into granodiorite whilst on the $2^*A/CNK-Na_2O/K_2O$ -2FMSB triangular plot (see Fig. 20b; Laurent et al., 2014) they plot on the anatectic granite field. On the K-Na-Ca diagram (see Fig. 20c) and Q-Ab-Or plot (see Fig. 20d; Barker and Arth, 1976) these granites follow classic calc-alkaline differentiation contrast to Archean TTGs displaying trondhjemite trend. Major element oxides on Harker binary diagrams define moderate to strongly fractionated trends against SiO_2 . Plots of TiO_2 , Al_2O_3 , Fe_2O_3 , MgO , CaO , Na_2O and P_2O_5 display negative trends whilst K_2O exhibit positive trends (Figure is not given). These potassic granite show affinity to A-type granite on the $(K_2O+Na_2O)/CaO$ versus $Zr+Nb+Ce+Y$ plot (Fig. 22b; Whalen et al., 1987) which is consistent with their high (Na_2O+K_2O) , high incompatible elements content (Rb, Ba, REE, Th, U, Nb, Zr, Y), lower CaO and Sr (Whalen et al., 1987; Bonin, 2007). Their high degree of magmatic differentiation (Figure is not given) together with Nd isotopic evidence for mantle input (Jayananda et al., 2019) also in conformity with their affinity to A-type granites. On the LREE versus Ba+Sr plot (Laurent et al., 2014) they plot on the biotite-granite field with few samples extending into sanukitoid or TTGs field (see Fig. 22a). The total REE range from 85.99 - 826.41 ppm and show poorly to strongly fractionated REE patterns with $[(La/Yb)_N = 2.74$ - $50.39]$ and negative Eu ($Eu/Eu^* = 0.26 - 0.67$) anomalies (Fig. 23a). They show strong negative Ba, Nb-Ta, Sr, Ti anomalies on the primitive mantle normalized multi-element diagrams whilst Y show positive anomalies (see Fig. 23a).

8.1.2. ca. 2600 Ma high-K granites

The ca. 2600 Ma old Chitradurga, Arsikere - Banavara, and Pandavpura plutons contain euhedral quartz crystals, coupled with K-feldspar, plagioclase (An_{10-16}) and euhedral magmatic epidote pointing to their rapid emplacement. They are characterized by higher SiO_2 , K_2O conversely lower Al_2O_3 compared to the ca. 3000 Ma granite plutons. On Ab-An-Or triangular diagram all the samples plot on granite field (see Fig. 20a) whilst on the $2^*A/CNK-Na_2O/K_2O$ -2FMSB plot (see Fig. 20b; Laurent et al., 2014) they fall on anatectic granite with few samples extending into TTG field. They define strong calc-alkaline differentiation trend on K-Na-Ca and Q-Ab-Or triangular plots (after Barker and Arth, 1976; see Figs. 20c-d). These potassic granites are characterized by high abundances of incompatible elements similar to 3000 Ma granite except lower Sr, Zr but conversely higher Nb and Y (figure not given). They contain moderate to very high total REE (85.57-749.65 ppm) and show poorly to moderately fractionated REE patterns $[(La/Yb)_N = 1.65$ - $22.98]$ with negative Eu ($Eu/Eu^* = 0.15 - 0.65$) anomalies (Fig. 23b). On the primitive mantle normalized multi-element plot these granites show strong negative Ba, Nb-Ta, Sr and Ti anomalies whilst positive or

no significant Y anomalies (see Fig. 23b)

8.2. Eastern corridor - ca. 3300-3180 Ma TTG-type gneisses

The TTG-type migmatitic gneisses to the east of the Chitradurga boundary shear zone show narrow range of major element compositions [$SiO_2 = 65.37$ to 73.82 wt.%; $Al_2O_3 = 14.28$ - 15.37 wt.%; $Na_2O = 4.35$ - 5.63 wt.%; $K_2O = 1.07$ - 2.05 wt.%; $MgO = 0.37$ - 1.42 wt.%]. on the Ab-An-Or triangular plot (O'Connor, 1965) they plot on tonalite and trondhjemite with few samples extending into granodiorite to granite (see Fig. 20a). They exhibit high Na_2O/K_2O ratios and plot overlapping TTG field on the $2^*A/CNK-Na_2O/K_2O$ -2FMSB triangular plot (see Fig. 20b; Laurent et al., 2014) and follow trondhjemite differentiation trend on K-Na-Ca and Q-Ab-Or plots (see Figs. 20c-d; after Barker and Arth, 1976). On the Harker binary plots, they follow the similar differentiation trends of western TTGs (figure not given). These migmatitic TTGs exhibit similar characteristics to western TTGs but show slightly lower contents of incompatible elements like Nb, Zr, Ba, Sr, Y and REE (35.33-202.58 ppm). They display poorly to strongly fractionated REE patterns $[(La/Yb)_N = 1.98$ - $85.30]$ without significant Eu anomalies ($Eu/Eu^* = 0.46 - 1.29$) anomalies except four samples showing either negative or positive Eu anomalies (Fig. 24a). On the primitive mantle-normalized multi-element diagrams the migmatitic TTGs show negative Ba, Nb-Ta and Ti anomalies except few samples show either negative or positive Y anomalies (see Fig. 24a).

8.2.1. ca. 2700-2600 Ma transitional TTGs

The transitional TTGs contain higher CaO , MgO , TiO_2 , K_2O and P_2O_5 compared to the 3500-3200 Ma western TTGs (Jayananda et al., 2020). They exhibit significant variation in major element compositions [$SiO_2 = 66.63$ to 73.80 wt.%; $Al_2O_3 = 13.08$ - 15.49 wt.%; $Na_2O = 2.66$ - 5.23 wt.%; $K_2O = 1.95$ - 3.56 wt.%; $MgO = 0.20$ - 2.17 wt.%]. On Ab-An-Or plot (O'Connor's, 1965) the transitional TTGs plot on tonalite, granodiorite and extending into granite and quartz-monzonite fields (see Fig. 20a) whilst plot on TTG extending to sanukitoids field on $2^*A/CNK-Na_2O/K_2O$ -2FMSB triangular plot (see Fig. 20b; Laurent et al., 2014). On K-Na-Ca plot and Q-Ab-Or plots (Barker and Arth, 1976) they define transitional from trondhjemite to calc-alkaline differentiation trend (see Fig. 20c-d). Plots of major element oxides against SiO_2 define moderate to strong differentiation trends on Harker's binary diagrams (Figure is not given) wherein Al_2O_3 , TiO_2 , Fe_2O_3 , MgO , CaO , P_2O_5 show negative correlation whilst K_2O show moderate positive trend. These rocks contain higher incompatible (Ba, Sr, Zr, Y) and compatible elements (Ni and Cr) compared to the TTGs of western and central blocks (Jayananda et al., 2020; present study).

The transitional TTGs exhibit wide range of total REE (29.31-325.17 ppm) and show poorly to highly fractionated REE patterns $[(La/Yb)_N = 2.76$ - $57.68]$ without significant Eu anomalies or positive Eu anomalies ($Eu/Eu^* = 0.73$ - 1.42) (Fig. 24b). On the primitive mantle normalized spider diagrams, they show strong negative Nb and Ti anomalies whilst negative to positive Sr, Ta, Nd, Sr and positive Y anomalies (see Fig. 24b).

8.2.2. ca. 2560-2530 Ma sanukitoids

Sanukitoids are characterized by high abundances of both incompatible and compatible elements (Moyen et al., 2001, 2003; Jayananda et al., 2020). These rocks show large variation in elemental compositions [$SiO_2 = 59.96$ to 71.10 wt.%; $Al_2O_3 = 14.08$ - 15.78 wt.%; $Na_2O = 3.58$ - 4.21 wt.%; $K_2O = 2.95$ - 4.68 wt.%; $MgO = 0.90$ - 3.61 wt.%]. On the Ab-An-Or triangular plot (O'Connor's, 1965) they plot on granodiorite with few samples extending into quartz-monzonite (see Fig. 20a) whilst plot on $2^*A/CNK-Na_2O/K_2O$ -2FMSB triangular plot falls on the sanukitoid field (see Fig. 20b; after Laurent et al., 2014). They follow calc-alkaline differentiation trend on K-Na-Ca diagram and Q-Ab-Or plot (Barker and Arth, 1976). Major oxides plotted against SiO_2 on the Harker binary diagrams (see Fig. 20c-d) define differentiation trends

wherein Al_2O_3 , TiO_2 , Fe_2O_3 , CaO , P_2O_5 show strong negative correlation whilst Na_2O display spread but K_2O show positive correlation (figure not given).

The sanukitoids exhibit high to very high content of incompatible elements (Ba, Sr, Zr, Nb, Y REE) and compatible elements (Ni, Cr). They show high to very high total REE (104.07–750.30 ppm) and display moderate to strongly fractionated REE patterns $[(\text{La}/\text{Yb})_N=7.51\text{--}47.23]$ with feeble negative or no significant Eu anomalies (Fig. 24c). On the primitive mantle normalized multi-element diagram they show prominent Nb-Ta, Hf and Ti anomalies with either positive or negative Y anomalies (see Fig. 24c).

8.2.3. ca. 2520–2513 Ma anatectic granites

The anatectic granites display significant variation in major element composition $[\text{SiO}_2=70.22 \text{ to } 76.32 \text{ wt.}\%; \text{Al}_2\text{O}_3=12.10\text{--}14.64 \text{ wt.}\%; \text{Na}_2\text{O}=3.11\text{--}3.86 \text{ wt.}\%; \text{K}_2\text{O}=3.49\text{--}5.68 \text{ wt.}\%; \text{MgO}=0.20\text{--}0.58 \text{ wt.}\%]$. All the samples of anatectic granites plot on the granite field of Ab-An-Or triangular (see Fig. 20a) diagram (O'Connor's, 1965) and 2*A/CNK- $\text{Na}_2\text{O}/\text{K}_2\text{O}$ -2FMSB triangular plot (see Fig. 20b; after Laurent et al., 2014). They follow calc-alkaline differentiation trend on K-Na-Ca and Q-Ab-Or triangular plots (see Figs. 20c–d; Barker and Arth, 1976). On the Harker's binary diagram, major oxides of Al_2O_3 , TiO_2 , Fe_2O_3 , CaO , P_2O_5 show negative correlation and K_2O exhibit positive trend. They show high content of Rb, Ba but low Sr and moderate to high total REE (62.71–467.30 ppm). Anatectic granites display poorly to highly fractionated REE patterns $[(\text{La}/\text{Yb})_N=3.06\text{--}81.01]$ with prominent negative Eu anomalies ($\text{Eu}/\text{Eu}^*=0.20\text{--}0.58$; Fig. 24d). On the Primitive mantle normalized multi-element plot, anatectic granites show strong negative Ba, Sr, Nb, Ti anomalies but variable Ta (negative and positive) and with positive or without any Y anomalies.

8.2.4. ca. 2500 Ma Maddakkanahalli granite

This fine-grained aplitic granite is the youngest Archean granite in the Dharwar craton found as sheets along flat to shallow foliation of transitional TTGs and Javagondanahalli greenstone belt. It contains high SiO_2 (72.69–76.80 wt.%), low to moderate Al_2O_3 (12.22–14.68 wt.%), very low MgO (0.0–0.03 wt.%) and high K_2O (4.10–6.48 wt.%). They plot on the granite field of Ab-An-Or triangular (see Fig. 20a) diagram (O'Connor's, 1965) and 2*A/CNK- $\text{Na}_2\text{O}/\text{K}_2\text{O}$ -2FMSB triangular plot (see Fig. 20b; after Laurent et al., 2014). The Maddakkanahalli granite contain high Rb, low Ba, Sr, low to high REE, Nb, Ta, Zr, Y. They display poorly to highly fractionated REE patterns $[(\text{La}/\text{Yb})_N=2.37\text{--}60.96]$ with negative Eu anomalies ($\text{Eu}/\text{Eu}^*=0.08\text{--}1.02$; Fig. 24e). Primitive mantle normalized spider diagram displays negative, Ba, Nb-Ta, Sr (one sample show positive) and Ti anomalies (see Fig. 24e).

9. Whole rock Nd isotopes

9.1. Greenstone volcanics

Sm-Nd isotope data of published record combined with our new Nd data for greenstone volcanics (Sargur Group and Dharwar Supergroup), ca. 3350–3300 Ma TTG-type gneisses, ca. 3000–2950 Ma and ca. 2600 Ma high-K granites from the western block as well as ca. 3200 Ma migmatitic TTGs, ca. 2690–2630 Ma transitional TTGs, ca. 2560–2530 Ma sanukitoids and ca. 2510 Ma anatectic granites are presented in Supplementary table 1.

The ca. 3400–3200 Sargur Group komatiites and komatiite volcanics from western margin of Chitradurga greenstone belt display Nd model ages of 3710–3430 Ma except a few samples show anomalous high or low values (Jayananda et al., 2008, 2023; Maya et al., 2017). The $\epsilon_{\text{Nd}(T)}$ values ranging from +3.0 to +1.3 with few anomalously high positive values up to +6.0 or few negative values ranging -1.0 to -2.5 (Jayananda et al., 2008). These values together with their low incompatible element contents can be attributed to their derivation from depleted mantle reservoirs with traces of possible ancient crustal contamination. The

studied Kibbanahalli arm belong to ca. 3000–2850 Ma Bababudan Group exhibit Nd model ages of 3383–3214 Ma (excluding anomalous values) with $\epsilon_{\text{Nd}(T)}$ values -0.2 to -1.5 suggesting their origin from primitive mantle sources or depleted mantle reservoirs with traces of ancient crustal contamination (Kumar et al., 1996; Jayananda et al., 2023; unpub. data). The ca. 2700 Ma Chitradurga Group volcanics show Nd model ages ranging 2732–3100 Ma with several samples displaying anomalously high values of ca. 3400–4000 Ma (present study; Jayananda et al., 2013b; Kumar et al., 1996; Rose et al., 2022). Their $\epsilon_{\text{Nd}(T)}$ values of +3.9 to -4.1 are in consistence with derivation from depleted mantle source with ancient crustal contamination. The ca. 2700 Ma Javagondanahalli mafic volcanics show Nd model ages of 3139–2835 Ma except few anomalous values >3200 Ma. Their $\epsilon_{\text{Nd}(T)}$ values [+3.8 to -0.4 except one sample displaying -2.3] indicate their origin from depleted to primitive mantle sources with traces of contamination pre-existing felsic crust (Fig. 25a).

9.2. Granitoids (TTGs, transitional TTGs, potassic granites, sanukitoids and anatectic granites)

The ca. 3350–3200 Ma old TTG-type gneisses in the western corridor forming the oldest basement show wide range of Nd model ages ranging 3664–3250 Ma with $\epsilon_{\text{Nd}(T)} = +2.9$ to -0.9. These values are consistent with their derivation from juvenile depleted sources and/or crustal protoliths with variable crustal residence periods. The ca. 3000–2950 Ma potassic granites display Nd model ages ranging from 3570 to 3070 Ma with $\epsilon_{\text{Nd}(T)} = -2.9$ to -0.2 except two samples exhibiting +1.9 to 0.2. These are values reveal involvement of pre-existing crustal sources with minor juvenile mantle input in their origin. The ca. 2600 Ma old potassic granites exhibit Nd model ages of 3357–2956 Ma with $\epsilon_{\text{Nd}(T)} = -6.2$ to -0.6 which argue for dominant ancient crustal source with minor juvenile input in their origin (see Fig. 25a).

East of Chitradurga boundary shear zone ca. 3300–3150 Ma old migmatitic gneisses exhibit Nd model ages ranging 3419–3249 Ma with $\epsilon_{\text{Nd}(T)} = +2.9$ to -0.8 consistent with their genesis by melting of juvenile sources with minor input of ancient crust. In contrast the 2690–2630 Ma transitional TTGs show younger Nd model ages ranging 2983–2697 Ma and $\epsilon_{\text{Nd}(T)} = +2.4$ to 0.7 except two samples displaying negative values (-2.6 to -1.5). These values are in consistent with their derivation from juvenile sources and minor ancient crustal input. The 2560–2540 Ma sanukitoids exhibit Nd model ages of 2995–2697 Ma with $\epsilon_{\text{Nd}(T)} = +2.0$ to +0.3 except three samples showing slight negative values (-0.9 to -1.6) suggesting major juvenile input and possible interaction of ascending magmas with pre-existing crust origin (see Fig. 25a). The ca. 2510–2507 Ma anatectic granite sheets found in the eastern basement including Maddakkanahalli granite (see map in Fig. 1b) display old model ages ranging 3400–2745 Ma (present study; Mohan et al., 2023) with negative $\epsilon_{\text{Nd}(T)}$ values of -4.9 to -2.4 except four samples showing 0 to +1. These values are in consistent with their derivation from composite sources involving mix of ancient crust (>3000 Ma) and young (<2700 Ma) juvenile crust (Jayananda et al., 2020).

10. LA-MC-ICM-PS Lu-Hf isotope measurements in zircons

Published record of in-situ zircon Lu-Hf analyses together with new results from the present work are presented in the Supplementary table 5 and Fig. 25b. The average 2SE precision obtained on measured $^{176}\text{Hf}/^{177}\text{Hf}$ is 0.000032 (i.e., 0.000021–0.000090; Table which is equivalent to an uncertainty of 1.1 ϵ_{Hf} unit (0.7–3.2).

10.1. Western corridor

Zircon crystals from 18 HOL 01B have $^{176}\text{Hf}/^{177}\text{Hf}$ ranging from 0.280825 to 0.280906 and correlated with both $^{176}\text{Lu}/^{177}\text{Hf}$ and $^{176}\text{Yb}/^{177}\text{Hf}$ which range from 0.0009 to 0.0011 and from 0.020 to 0.029, respectively. The average initial $^{176}\text{Hf}/^{177}\text{Hf}$ shown by analyzed

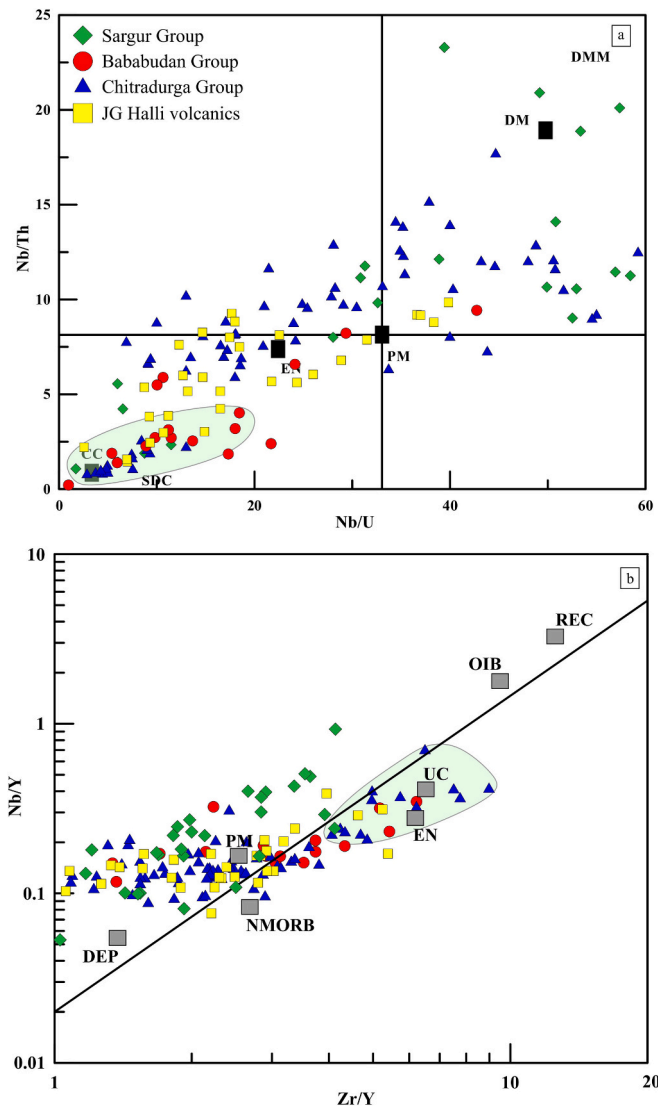


Fig. 26. (a) Nb/U versus Nb/Th plot (after Saunders et al., 1988) (b) Nb/Y versus Zr/Y (Condie, 2003).

zircon at 3222 Ma is 0.281017 ± 0.000051 (2SD) which translates into an ϵ_{Hf} of $+4.3 \pm 1.8$ (2SD).

Zircon crystals from 20 WCH 01A have $^{176}\text{Hf}/^{177}\text{Hf}$ that range between 0.280827 and 0.280872, whereas $^{176}\text{Lu}/^{177}\text{Hf}$ varies from 0.0007 to 0.0012 and $^{176}\text{Yb}/^{177}\text{Hf}$ from 0.019 to 0.033. The initial $^{176}\text{Hf}/^{177}\text{Hf}$ at 3237 Ma shown by analyzed zircons are consistent with each other and give an average of 0.280790 ± 0.000037 (2SD) which translates into an ϵ_{Hf} of $+3.5 \pm 1.3$ (2SD).

Zircon crystals from 20 WCH 06C have $^{176}\text{Hf}/^{177}\text{Hf}$ ranging from 0.280833 to 0.280958, $^{176}\text{Lu}/^{177}\text{Hf}$ from 0.0006 to 0.002, and $^{176}\text{Yb}/^{177}\text{Hf}$ from 0.016 to 0.063. The initial $^{176}\text{Hf}/^{177}\text{Hf}$ at 3229 Ma are broadly consistent and result in an average of 0.280807 ± 0.000055 (2SD) which is equivalent to an ϵ_{Hf} of $+4.1 \pm 2.0$ (2SD).

Zircon crystals from 20 WCH 06A have $^{176}\text{Hf}/^{177}\text{Hf}$ ranging from 0.280766 to 0.280959, $^{176}\text{Lu}/^{177}\text{Hf}$ from 0.0001 to 0.0005, and $^{176}\text{Yb}/^{177}\text{Hf}$ from 0.003 to 0.012. The initial $^{176}\text{Hf}/^{177}\text{Hf}$ at 3209 Ma are variable and result in an average of 0.280870 ± 0.000163 (2SD) which is equivalent to an ϵ_{Hf} of $+5.9 \pm 5.8$ (2SD).

Zircon crystals from 20 WCH 07A have $^{176}\text{Hf}/^{177}\text{Hf}$ that range between 0.280854 and 0.280949, whereas $^{176}\text{Lu}/^{177}\text{Hf}$ varies from 0.0008 to 0.0015 and $^{176}\text{Yb}/^{177}\text{Hf}$ from 0.025 to 0.043. The calculated initial $^{176}\text{Hf}/^{177}\text{Hf}$ at 3234 Ma give an average of 0.280836 ± 0.000098 (2SD)

which translates into an ϵ_{Hf} of $+5.2 \pm 3.5$ (2SD).

Zircons from 20 WCH 5B Hosdurga granite sample have $^{176}\text{Hf}/^{177}\text{Hf}$ values ranging from 0.280943 to 0.280999 with $^{176}\text{Lu}/^{177}\text{Hf}$ values varies between 0.01313 to 0.2263 and $^{176}\text{Yb}/^{177}\text{Hf}$ values ranging 0.401 to 0.0689. The calculated initial ratio of 2954.8 Ma of 0.280875 ± 0.000064 which corresponds to ϵ_{Hf} of $+0.0 \pm 2.3$ (2SD).

Zircon crystals from potassic granite sample BC 01 have $^{176}\text{Hf}/^{177}\text{Hf}$ that range between 0.280886 and 0.281026, whereas $^{176}\text{Lu}/^{177}\text{Hf}$ varies from 0.0003 to 0.0016 and $^{176}\text{Yb}/^{177}\text{Hf}$ from 0.005 to 0.054. The calculated initial $^{176}\text{Hf}/^{177}\text{Hf}$ at 2977 Ma give a relatively precise average of 0.280897 ± 0.000049 (2SD) which translates into an ϵ_{Hf} of $+1.3 \pm 1.7$ (2SD).

10.2. Eastern corridor

Zircon crystals from granodioritic TTG sample 19AL01A exhibit $^{176}\text{Hf}/^{177}\text{Hf}$ values ranging 0.280753 to 0.281006 whereas $^{176}\text{Lu}/^{177}\text{Hf}$ data range 0.000942 to 0.02446 and $^{176}\text{Yb}/^{177}\text{Hf}$ range 0.0120 – 0.1183. The calculated initial $^{176}\text{Hf}/^{177}\text{Hf}$ value at 3338 ± 17 Ma is 0.280762 ± 0.000124 which is equivalent to a ϵ_{Hf} value of $+5.1 \pm 4.4$ (2SD).

Tonalite sample 20AL01A show in-situ zircon $^{176}\text{Hf}/^{177}\text{Hf}$ values between 0.280739 to 0.280884 whilst $^{176}\text{Lu}/^{177}\text{Hf}$ values vary from 0.000942 to 0.002446 and $^{176}\text{Yb}/^{177}\text{Hf}$ range 0.0201 to 0.0746. The initial $^{176}\text{Hf}/^{177}\text{Hf}$ value at 3258 ± 8 Ma is 0.280743 ± 0.000115 which corresponds to a ϵ_{Hf} value of $+2.5 \pm 4.1$ (2SD).

Zircon crystals from sample 18CDA2 (Transitional TTG) exhibit $^{176}\text{Hf}/^{177}\text{Hf}$ values ranging from 0.281198 to 0.281302 with $^{176}\text{Lu}/^{177}\text{Hf}$ values vary 0.000397–0.001491 and $^{176}\text{Yb}/^{177}\text{Hf}$ range of 0.0108–0.0448. The calculated initial $^{176}\text{Hf}/^{177}\text{Hf}$ at 2633 ± 9 Ma is 0.281183 ± 0.000081 which translates into ϵ_{Hf} value of $+3.3 \pm 2.9$ (2SD).

Zircon crystals from Transitional TTG sample 18CD19 show in-situ $^{176}\text{Hf}/^{177}\text{Hf}$ values ranging from 0.281179 to 0.281239, $^{176}\text{Lu}/^{177}\text{Hf}$ values vary between 0.000488 – 0.000660 and $^{176}\text{Yb}/^{177}\text{Hf}$ values ranging 0.0136 – 0.0189. The calculated initial $^{176}\text{Hf}/^{177}\text{Hf}$ value 0.281185 ± 0.000046 at 2647 ± 17 Ma which corresponds to a ϵ_{Hf} value of $+3.7 \pm 1.6$ (2SD).

Zircon crystals from Transitional TTGs sample 18CD14 have in-situ $^{176}\text{Hf}/^{177}\text{Hf}$ values ranging from 0.281087 to 0.281200 with $^{176}\text{Lu}/^{177}\text{Hf}$ values between 0.000964–0.001546 and $^{176}\text{Yb}/^{177}\text{Hf}$ values ranging 0.0251– 0.0402. The initial $^{176}\text{Hf}/^{177}\text{Hf}$ value 0.281086 ± 0.000118 at 2640 ± 14 Ma translate into ϵ_{Hf} value of $+0.2 \pm 4.2$ (2SD).

The two analyses done on the only zircon recovered from Transitional TTG sample RAV 01 gave $^{176}\text{Hf}/^{177}\text{Hf}$ of 0.281156 ± 0.000034 and 0.281234 ± 0.000036 , which correlate with $^{176}\text{Lu}/^{177}\text{Hf}$ (0.00097 and 0.00143) and $^{176}\text{Yb}/^{177}\text{Hf}$ (0.0318 and 0.0483). The calculated initial $^{176}\text{Hf}/^{177}\text{Hf}$ at 2618 ± 9 Ma gives an average of 0.281134 ± 0.000078 (2SD) which translates into an ϵ_{Hf} of $+1.2 \pm 2.8$ (2SD).

Zircon crystals from Transitional TTG sample 17CL01 show in-situ $^{176}\text{Hf}/^{177}\text{Hf}$ values ranging from 0.281171 to 0.281306 with $^{176}\text{Lu}/^{177}\text{Hf}$ values between 0.000314–0.001351 and $^{176}\text{Yb}/^{177}\text{Hf}$ values ranging 0.0096– 0.0389. The initial $^{176}\text{Hf}/^{177}\text{Hf}$ value 0.281067 ± 0.000070 at 2669 ± 7.7 Ma corresponds to a ϵ_{Hf} value of $+5.1 \pm 2.5$ (2SD).

Zircon crystals from J.N. Kote granite sample 17JNK01 show in-situ $^{176}\text{Hf}/^{177}\text{Hf}$ values ranging from 0.281038 to 0.281132, $^{176}\text{Lu}/^{177}\text{Hf}$ values vary between 0.000774 – 0.002594 and $^{176}\text{Yb}/^{177}\text{Hf}$ values ranging 0.0216 – 0.0800. The calculated initial $^{176}\text{Hf}/^{177}\text{Hf}$ value 0.281014 ± 0.000023 at 2597 ± 11 Ma which corresponds to a ϵ_{Hf} value of -3.5 ± 0.8 (2SD).

Zircon crystals from granite sample 18CD21 have in-situ $^{176}\text{Hf}/^{177}\text{Hf}$ values ranging from 0.280916 to 0.281165 with $^{176}\text{Lu}/^{177}\text{Hf}$ values between 0.000390–0.001631 and $^{176}\text{Yb}/^{177}\text{Hf}$ values ranging 0.0120– 0.0476. The initial $^{176}\text{Hf}/^{177}\text{Hf}$ value 0.281067 ± 0.000232 at 2537 ± 8 Ma translate into ϵ_{Hf} value of -3.1 ± 8.1 (2SD).

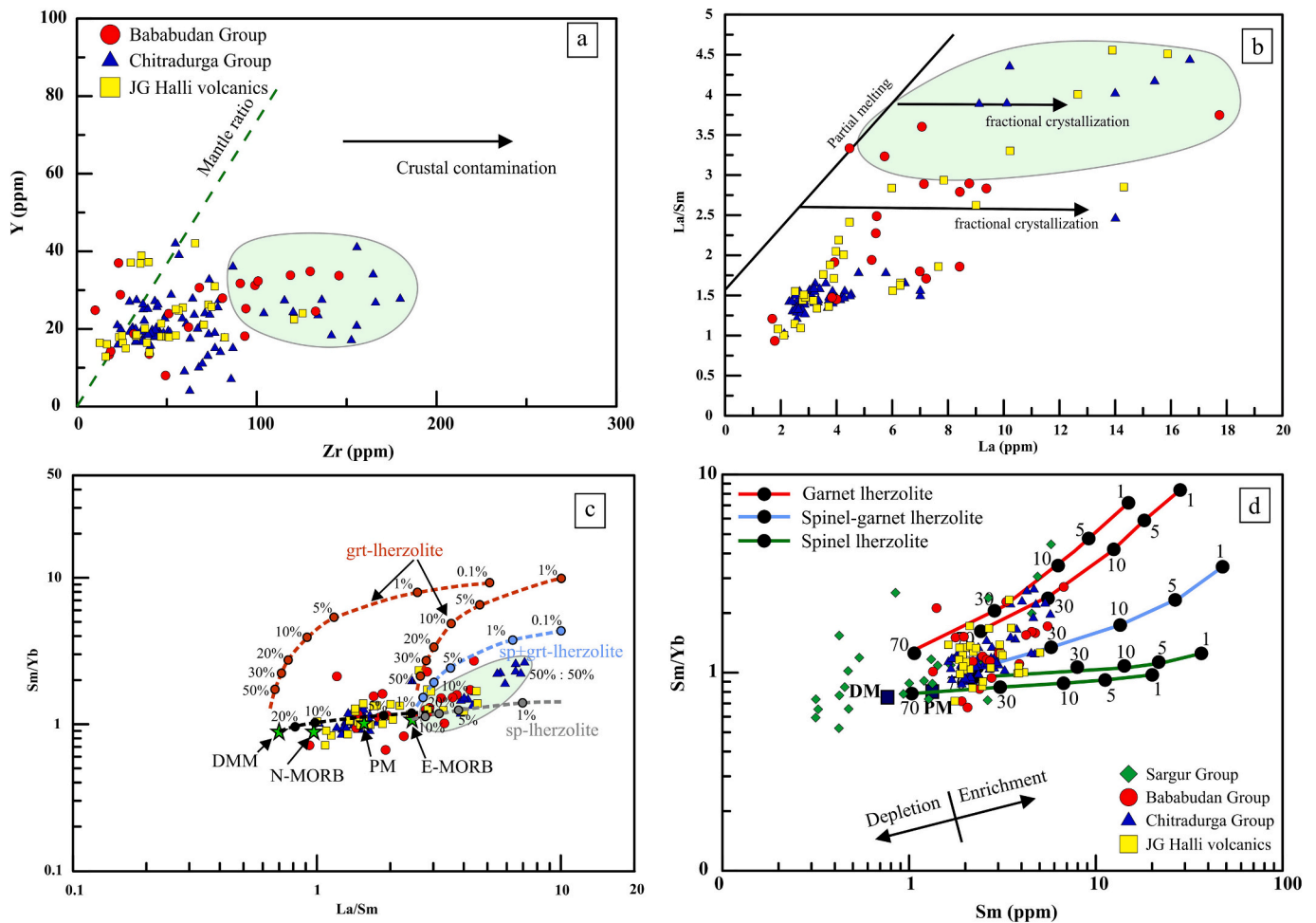


Fig. 27. (a) Y (ppm) versus Zr (ppm) binary diagram for Bababudan, Chitradurga and Javagondanahalli greenstone volcanics (b) La (ppm) versus La/Sm plot (after Treuil and Joron, 1975) for the Bababudan, Chitradurga and Javagondanahalli volcanics (c) La/Sm versus Sm/Yb binary plot (Aldanmaz et al., 2000) indicate source reservoirs from depleted to enriched mantle composition through primitive mantle at shallow depth with 5 to 10% melting of spl+lherzolite (d) Sm/Yb versus Sm binary plot (Aldanmaz et al., 2000) indicating the melting of source reservoirs sp-lherzolite to sp+grt lherzolite.

11. Petrogenesis of greenstone volcanic and granitoids

11.1. Effects of secondary processes on the elemental and isotope compositions

Major and trace element data together with radiogenic and stable isotope data for Archean greenstone-granitoid assemblages reveal mobility of some elements during fluid induced hydrothermal alteration processes and greenschist to amphibolite facies metamorphism (Mathieu, 2018; Lahaye et al., 1995; Polat et al., 2002; Polat and Hofmann, 2003; Chavagnac, 2004; Jayananda et al., 2016, 2023). The greenstone volcanic rocks of the studied Dharwar corridor experienced greenschist to lower amphibolite facies metamorphism. These volcanics were also affected by fluid flow as revealed by tiny network of carbonate veins which traverse the ultramafic volcanics. The primary mineralogy of the volcanics is nearly absent except for rare remnants of olivine and orthopyroxene (enstatite) preserved in komatiites from Nuggihalli and J.C. Pura belts. From published record and new data of present study we have selected samples with low Loss on Ignition. Jayananda et al. (2008, 2016, 2018, 2020, 2023) revealed that secondary processes affected Archean greenstone volcanic and granitoid LIL elements and rarely Ce contents but no significant effect on the High Field Strength (HFS) elements including REE were noticed. Major element oxides (except alkalis) together with uniform $\text{CaO}/\text{Al}_2\text{O}_3$ and $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratios and trace elements (Ni, Cr, Y, Sr, Zr) of greenstone volcanics show moderate to

strong linear trends in binary diagrams which preclude major effects of alteration on these elemental concentrations. Furthermore, most of the samples show smooth REE patterns and primitive mantle normalized multi-element diagrams (except rare Ce anomalies) which also suggest that trace element contents were not significantly affected by secondary processes. The $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ values of greenstones and granitoids define linear arrays except for a few samples which could be due to re-opening of Sm-Nd isotope system during intense flow where the fluid/rock ratio was high which is consistent with the large scattering of LIL elements. Therefore, to minimize the effects of alteration on the chemistry of the greenstone volcanics and granitoids, we used the least mobile elements for petrogenetic discussions that follow.

11.2. Origin of greenstone volcanics – Western corridor

11.2.1. Sargur Group komatiite and komatiite basalts

Several focused studies discussed the origin of Archean komatiites including their potential mantle source reservoirs, thermal and chemical evolution of deep mantle source reservoirs, depth of melting as well as tectonic context of magma generation and their spatial link to continental growth in the early Earth (e.g., Arndt et al., 2008; Herzberg, 1995; Parman et al., 2001; Chavagnac, 2004; Mole et al., 2015; Kusky et al., 2018; McKenzie, 2020; Jayananda et al., 2023).

The komatiite-komatiitic basalt volcanic sequences in Ghattihosahalli, J.C. Pura, Banasandra and Nagamangala in the western corridor

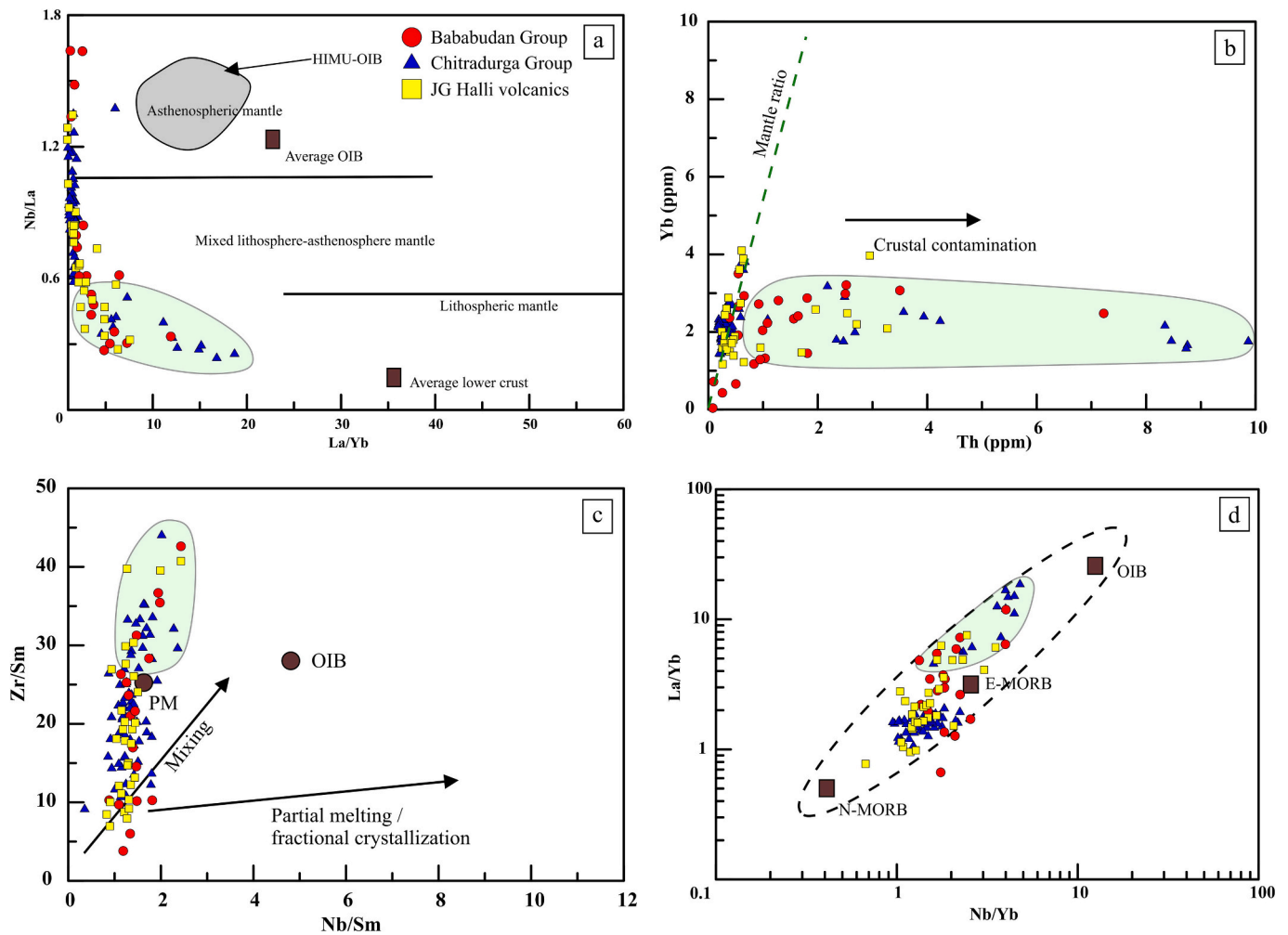


Fig. 28. (a) Nb/La versus La/Yb plot (after Abdel et al., 2004) for the Bababudan, Chitradurga and Javagondanahalli greenstone belts (b) Yb (ppm) versus Th (ppm) plot for Bababudan, Chitradurga and Javagondanahalli volcanics (c) Zr/Sm versus Nb/Sm ratio plot (after Bai et al., 2022) for the Bababudan, Chitradurga and Javagondanahalli greenstone volcanics (d) La/Yb versus Nb/Yb plot (after Pearce, 2008) for the Bababudan, Chitradurga and Javagondanahalli greenstone volcanics.

contain abundant pillow lavas and also display spinifex structures which unambiguously point to their eruption in a marine environment away from sizable continental landmass. This is largely consistent with elemental (high MgO, Ni, Cr, very low Rb, Ba, Sr, REE) and Nd isotope data [$\epsilon_{Nd(T)} = +4.0$ to 0.0 ; Jayananda et al., 2023; present study] which preclude any significant contamination from pre-existing continental crust. Therefore, field evidence together with elemental and Nd isotope data suggest komatiitic magmas originated beneath the oceanic lithosphere far from sizable pre-existing continental land mass.

The common differentiation trends defined by komatiites, komatiitic basalts (along with five basalt samples) on the Harker's binary diagrams indicate that they evolved through differentiation from ultramafic parental magma. The samples with very high MgO contents (>36.0 wt %) could possibly correspond to cumulates. The high MgO contents (15.27–42.41 wt.% excluding five basaltic samples) of komatiites coupled with very low to moderate total REE (1.30–22.51 ppm except few samples showing >45.0 ppm) with nearly flat REE patterns indicate that their parental magmas originated from high degree (40–45%) melting of mantle (Herzberg, 1992; Sossi et al., 2016) at great depth where melts were in equilibrium with the residual ultramafic mantle. Elemental characteristics particularly Al-depleted to Al-undepleted character [low to high $\text{CaO}/\text{Al}_2\text{O}_3$ (2.50–1.03 with very few anomalous values of <0.5), $\text{Al}_2\text{O}_3/\text{TiO}_2$ (6.7–17.1 with few anomalous values >30), $(\text{Gd}/\text{Yb})_N = 4.38\text{--}0.19$], combined with negative to positive Zr, Hf and Y anomalies argue for their origin in the deep mantle at depths

ranging from 400 to 250 km with variable residual garnet or without any garnet in their source residue (Jayananda et al., 2008, 2016, 2023; Maya et al., 2017). Nd isotopes ($\epsilon_{Nd(T)} = 0$ to $+4$, except a few anomalous values showing high positive values; see Fig. 25a) coupled with variable incompatible element contents (REE, Zr, Y, Nb, Ta, Th) and elemental ratios (Nb/Th vs Nb/U; Fig. 26a; after Saunders et al., 1988) reveal their origin from heterogeneous source reservoirs ranging from deep depleted to primitive mantle reservoirs. Further, more incompatible element ratios of studied komatiites and komatiitic basalts are also consistent with their derivation from primitive to highly depleted mantle as reflected by their Nb/Y vs Zr/Y plots (Fig. 26b; after Condie, 2003).

Elemental characteristics like high MgO contents coupled with low incompatible element abundances, phase equilibrium modelling and estimated magma temperatures (1700 to 1900°C) argue for the origin of komatiitic magmas by dry melting of mantle at great depths (5–10 Gpa) and in hot-spot environments (e.g. Herzberg, 1992; Campbell et al., 1989; Chavagnac, 2004; Arndt, 2008; Jayananda et al., 2016, 2023; Sossi et al., 2016). In contrast, Sobolev et al. (2016, 2019) proposed the origin of Archean komatiites by melting of deep hydrous mantle reservoirs associated with mantle plume whilst few other workers argued that komatiitic magmas can be generated by melting of shallow hydrous mantle in subduction zone (Parman et al., 2001; Ma et al., 2024).

Incompatible element ratios like Zr/Nb versus Nb/Th plot (Figure not given after Condie, 2003) are used to discriminate tectonic settings of ultramafic to mafic volcanics where most of the studied

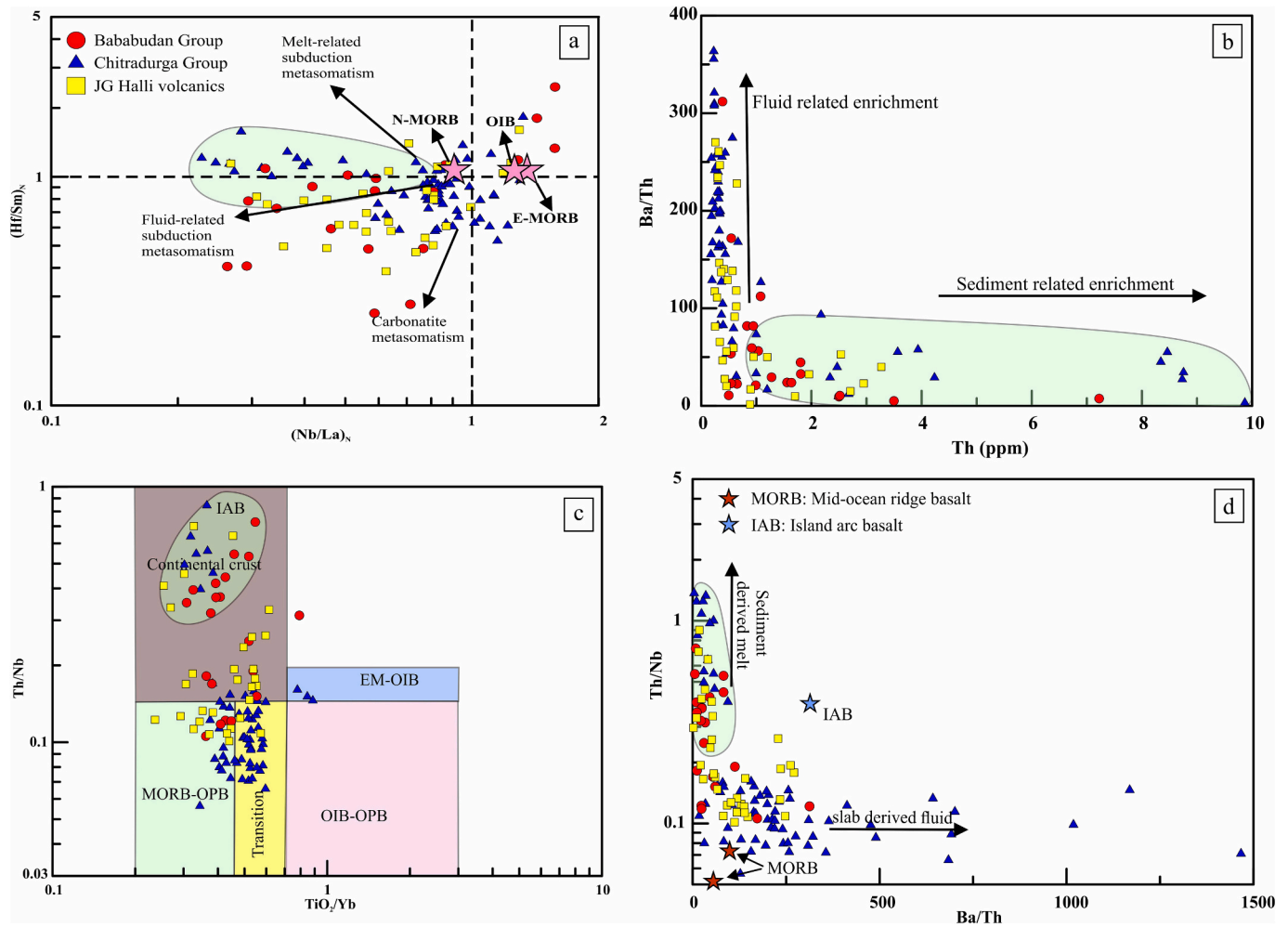


Fig. 29. (a) $(\text{Hf}/\text{Sm})_N$ versus $(\text{Nb}/\text{La})_N$ (after Hu et al., 2021) (b) Ba/Th versus Th (after La Fleche et al., 1998) suggest that mantle sources in the arc environment were enriched by slab derived fluids and/or interaction of sediment derived melts with arc mantle in subduction zone (c) Th/Nb vs TiO_2/Yb (after Pearce et al., 2021) for Bababudan, Chitradurga and Javagondanahalli mafic volcanics (d) Th/Nb vs Ba/Th plot (Saccani et al., 2018) depicts involvement of both sediment derived melts and subducting slab derived fluids in the enrichment of mantle.

komatiites and komatiitic basalts plot in oceanic plateau whilst few samples fall either on arc to N-MORB field which is also consistent with published results further southwest in the Holenarsipur region (Jayananda et al., 2023). This further implies that Paleoproterozoic (ca. 3380–3200 Ma) komatiite-komatiitic basalts from the western block were generated in hotspot environments and erupted as volcanic plateaus with few komatiitic basalt-basalts erupted as oceanic arc adjacent to the plateau (Jayananda et al., 2023). Few komatiite samples displaying highly depleted characteristics (total REE <5.0 ppm) might represent preserved fragments of oceanic ridge which accreted to an arc during convergence close to ca. 3200 Ma. Isotopic age data indicate that the time span between the formation of this oceanic plateau and subduction related ultramafic-mafic to felsic volcanic arc was about 50 Ma (Peucat et al., 1995; Jayananda et al., 2023). To summarize, the studied Sargur Group ultramafic komatiite-komatiitic basalt assemblages of the studied corridor in the western block originated from primitive mantle to deep depleted mantle reservoirs and emplaced as volcanic plateaus with minor oceanic arc or preserved ridge components.

11.2.2. Bababudan Group mafic volcanics (Kibbanahalli arm and Yediyur belt)

Field data such as vesicular to amygdaloidal textures (see Fig. 3a) together with rare angular gneiss enclaves in the mafic volcanics suggest their eruption largely on pre-existing continental crust which is also

reflected on Y vs Zr diagram (Fig. 27a after Palme and O'Neill, 2014) depicting crustal contamination. Although they show significant variation in SiO_2 contents (47.33–53.66 wt.%), large scattering of major element concentrations on Harker binary plots (Figure is not given) preclude any major differentiation of their parental magmas which is consistent with trace element ratios (La/Sm vs La plot; Fig. 27b after Treuil and Joron, 1975) depicting mainly partial melting trend with few samples following fractional crystallization. The wide range of total REE contents (28.27–158.41 ppm) combined with flat to moderately fractionated REE patterns [$(\text{Gd}/\text{Yb})_N = 0.87\text{--}2.09$; see Fig. 19a] suggest their origin by different degree melting of heterogeneous mantle reservoirs at different depths with variable or without any garnet in source residue. Niobium (Nb) anomalies in ultramafic-mafic volcanics are used to discriminate arc from plume sources as Nb behaves as a compatible element in shallow arc mantle where it is accommodated in residual ilmenite or rutile whilst in deeper mantle it behaves as incompatible and preferentially enters into melts. Therefore, negative Nb anomalies are attributed to magma origin in an arc setting whilst positive anomalies are explained by a plume source. Significant negative Nb anomalies with variable Zr, Hf, Ti and Y (negative with few positive) anomalies on primitive mantle normalized multi-element diagrams are also consistent with the origin of magmas from different depths with or without residual garnet. Large variations in MgO (5–19 wt.%) and SiO_2 (47–64 wt.%) contents coupled with incompatible element ratios (Sm/Yb versus $\text{La}/$

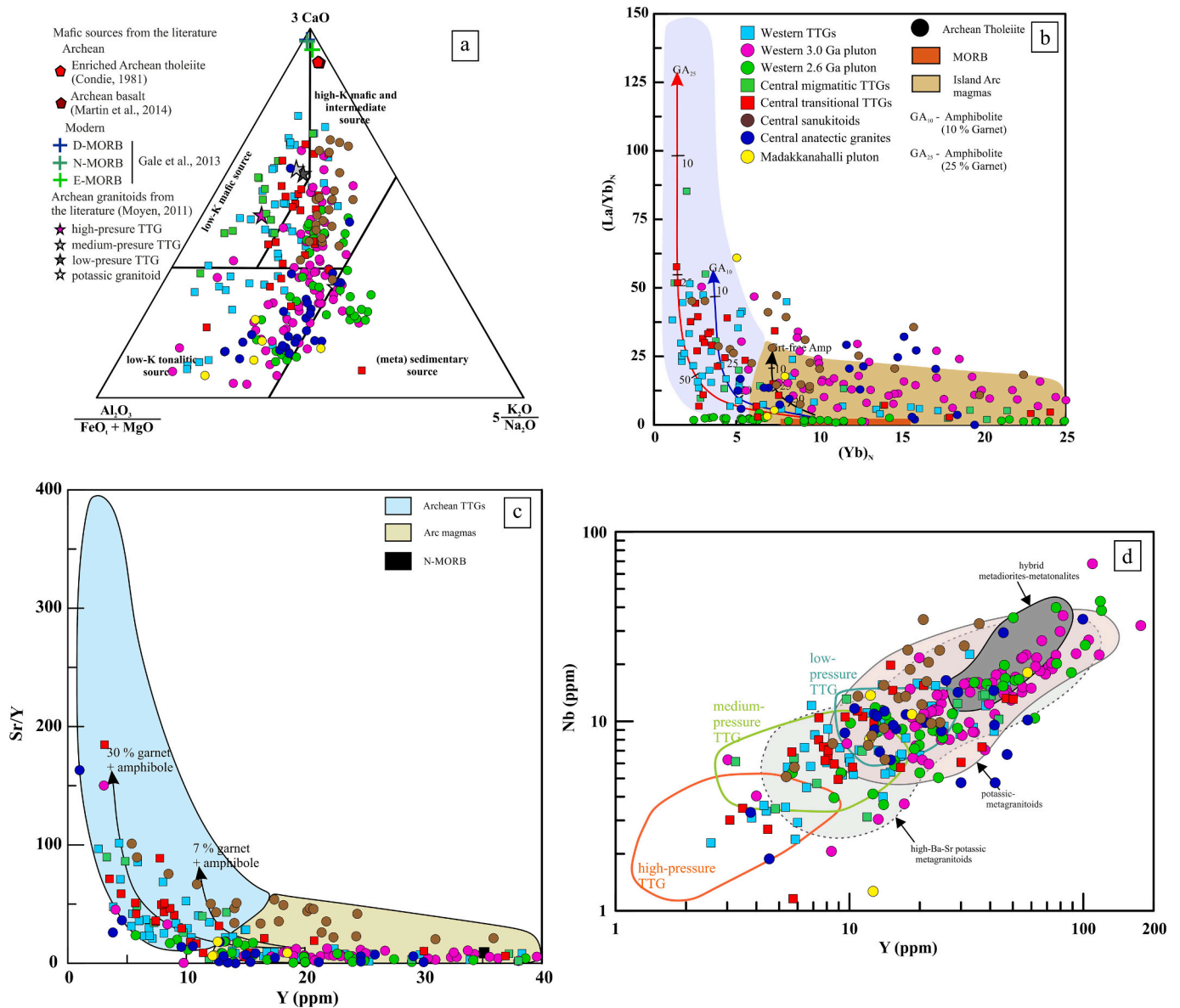


Fig. 30. (a) $\text{Al}_2\text{O}_3/(\text{FeO} + \text{MgO}) \cdot 3\text{CaO} \cdot 5\frac{\text{K}_2\text{O}}{\text{Na}_2\text{O}}$ diagram (Laurent et al., 2014) depicting sources of western and eastern basement granitoids (b) $(\text{La}/\text{Yb})_N$ versus $(\text{Yb})_N$ plot (after Jahn et al., 1981; Drummond and Defant, 1990; Moyen, 2011) showing the characteristic of residue garnet and amphibole and arrows indicate batch of melting of mafic source at deeper levels transforming into garnet bearing amphibolites residue with 7–30% garnet in residue (c) Sr/Y versus Y (Martin, 1986), reveal melting sources at different depths with variable garnet (7–30%) and/or amphibole in residue (d) Nb versus Y (after Simon et al., 2021) depicting medium to high pressure conditions.

Sm; Aldanmaz et al., 2000; Fig. 27c) argue for 20–30% melting of primitive mantle to E-MORB source in garnet-lherzolite to spinel lherzolite stability field which is in agreement with Sm/Yb versus Sm plot (Fig. 27d after Aldanmaz et al., 2000). Other elemental ratios such as $[\text{Sm}/\text{Yb}]$ versus Sm ; see Fig. 27d) also useful to discriminate potential mantle source reservoirs, degree and depths of melting (Aldanmaz et al., 2000) also favour origin of the Bababudan Group mafic volcanics from primitive mantle at different depths within stability fields of spinel-lherzolite to garnet-lherzolite. Incompatible element ratios (Nb/La vs La/Yb ; Fig. 28a, after Abdel and Nassar, 2004) used to constrain magma source reservoirs particularly evaluate lithospheric and asthenosphere mantle inputs where Bababudan mafic volcanics follow the mixing trend of lithospheric to asthenospheric mantle reservoirs which suggest their origin from different depths in the mantle. Nb/Th versus Nb/U plot (see Fig. 26a, Saunders et al., 1988) are used to constrain compositions of potential mantle source reservoirs where Bababudan volcanics reveal

their origin mainly from primitive mantle ($\text{Nb}/\text{Th}=8$; Condie, 2005) with significant crustal contamination as reflected by their low Nb/Th values (<8.0). This is consistent with other incompatible element ratios like (Th versus Yb plot; Fig. 28b) and Zr/Sm versus Nb/Sm ratios (c.f., Bai et al., 2022; Fig. 28c).

The $\text{Sm}-\text{Nd}$ isotope system is generally not much affected by post-magmatic low grade- metamorphism, fluid flow and hydrothermal alteration. Nd isotope compositions of mafic volcanics particularly high Nd model ages (3214–3572 Ma) compared to their whole rock isochron ages (2911–2850 Ma; Kumar et al., 1996) reveal possibility of contamination of pre-existing granitoid basement. Further, slight negative ϵ_{Nd} (τ) values (-0.2 to -1.5) coupled with incompatible element contents (REE , Nb , Zr , Y , U , Th) suggest their derivation from primitive mantle reservoirs with minor crustal contamination.

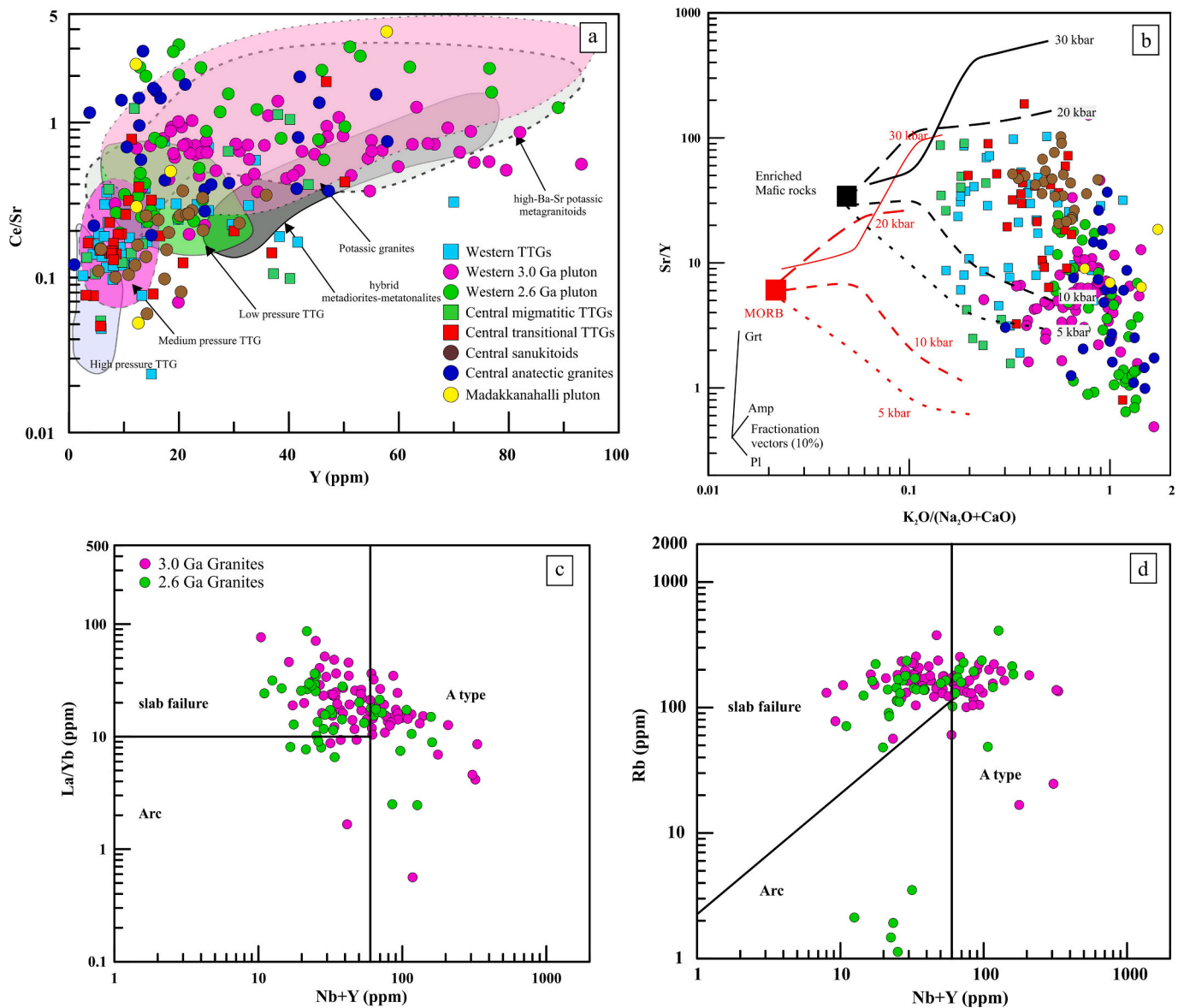


Fig. 31. (a) Ce/Sr versus Y (after [Moyen, 2011](#)) for western TTGs, eastern migmatitic, transitional TTGs, sanukitoids and anatectic granites (b) Sr/Y versus $K_2O/(Na_2O+CaO)$ plot (Hoffmann et al., 2019) indicating TTG magmas generated by melting of heterogeneous mafic source at medium to high pressure 8–20 kb pressure conditions (c) La/Yb versus $Nb+Y$ (after Whalen and Hilderbrand, 2019) and (d) Rb versus $Nb+Y$ (after [Pearce et al., 1984](#)) for potassic plutons suggest geodynamic setting from slab failure associated with collision of arcs.

11.2.3. Chitradurga volcanics

Elemental characteristics such as wide range of major element compositions ($SiO_2 = 44.81\text{--}61.17$ wt.%; $MgO = 4.70\text{--}11.48$ wt.%) coupled with large spread on major and trace elements on Harker binary plots preclude large scale magmatic differentiation processes of parental magmas. Alternatively different melting degrees of source reservoirs at varying depths in the mantle can be considered. This is clearly reflected in a La/Sm versus La diagram (see [Fig. 27b](#) after [Treuil and Joron, 1975](#)) where the majority of samples follow the partial melting trend except for few samples which scatter on fractional crystallization paths. Furthermore, minor crustal contamination cannot be totally ruled out because of negative $\epsilon_{Nd(T)}$ values of several samples ([Jayananda et al., 2013a](#); this study). The diverse incompatible element contents of mafic volcanic rocks including low total REE (25.58–48.21 ppm) to high total REE (54.54–145.54 ppm) together with large variations in abundances of other incompatible elements ($Ba = 432\text{--}21$ ppm; $Sr = 254\text{--}39$ ppm; $Th = 8.75\text{--}0.31$ ppm; $U = 1.48\text{--}0.08$ ppm; $Zr = 180\text{--}20$ ppm; $Hf = 4.4\text{--}0.94$ ppm; $Y = 39\text{--}16$ ppm) suggest their derivation from heterogeneous

mantle source reservoirs. Mafic volcanics with low total REE (<50.0 ppm) coupled with flat REE patterns [$(Gd/Yb)_N = <1.0$] $0.92\text{--}1.92$; $(La/Yb)_N = <1.5$] $0.70\text{--}12.35$] point to higher degree of melting of shallow depleted mantle. In contrast, mafic volcanics with higher total REE (>100 ppm), coupled with higher Rb , Ba , Sr , Th , U , Zr , Hf and Y suggest possible undepleted to enriched sources with or without crustal contamination. The high SiO_2 and high incompatible element contents of the intermediate to felsic pyroclastic volcanics infer remelting of newly formed arc crust with significant pre-existing crustal input. Incompatible element ratios have been extensively used to characterize the composition and evolution of mantle source reservoirs ([Aldanmaz et al., 2000](#); [Pearce, 2008](#); [Saunders et al., 1988](#); [Shuto et al., 2006](#); [Saccani et al., 2017](#); [Hu et al., 2021](#)). The Nb/Th versus Nb/U plot (see [Fig. 26a](#), after [Saunders et al., 1988](#)) reveal the origin of Chitradurga volcanics from heterogeneous mantle sources ranging in composition from depleted mantle, primitive to enriched mantle with minor crustal input which is consistent with Yb versus Th plot (see [Fig. 28b](#) after [Manikyamba et al., 2022](#)). Several other incompatible element ratios

such as La/Yb versus Nb/Yb (Fig. 28d after Pearce, 2008), Nb/Y versus Zr/Y (see Fig. 26b Condie, 2003) and Sm/Yb versus La/Sm (see Fig. 27c after Aldanmaz et al., 2000) are also consistent with heterogeneous mantle reservoirs that correspond to depleted, primitive (E-MORB) and enriched mantle with minor crustal contamination. The incompatible element data is consistent with Nd isotope data ($\epsilon_{\text{Nd}(T)} = +3.9$ to $+1.1$ and -1.1 to -4.1) indicating heterogeneous source reservoirs such as primitive to enriched mantle and depleted mantle, and variable contamination of pre-existing crust. However, chemical evolution of their heterogeneous source reservoirs is not clear. Globally, the origin of depleted mantle reservoirs is attributed to episodic extraction of the continental crust from the mantle (e.g., Stein and Hofmann, 1994; Boyet and Carlson, 2005). The observed depleted signatures in some mafic volcanics (low incompatible elemental contents and Nd [$\epsilon_{\text{Nd}(T)} > +2$]) reveal depleted mantle sources which can be attributed to the extraction of crust in earlier episode probably prior to ca. 3000 Ma, as revealed by a major crust forming event that contributed to continental growth during

ca. 3200–3100 Ma (Jayananda et al., 2013b; Mohan et al., 2023; Ravindran et al., 2023) indicating long term depletion of the mantle. However, processes involved in the evolution of enriched source reservoirs is not clear. Several studies on the Neoproterozoic crust accretion shown that their mantle sources were enriched either through fluids from the down going slab or melts produced by melting of down-going slab with sediments (e.g., La Flèche et al., 1998; Hu et al., 2021). Incompatible element ratios like (Hf/Sm)_N versus (Nb/La)_N (Fig. 29a, Hu et al., 2021) and Ba/Th versus Th (Fig. 29b, La Flèche et al., 1998) suggest that mantle sources in the arc environment were enriched by slab derived fluids and/or interaction of sediment-derived melts with arc mantle in subduction zone. In summary, the Chitradurga volcanics originated from heterogeneous mantle source reservoirs (depleted to enriched) with minor crustal contamination.

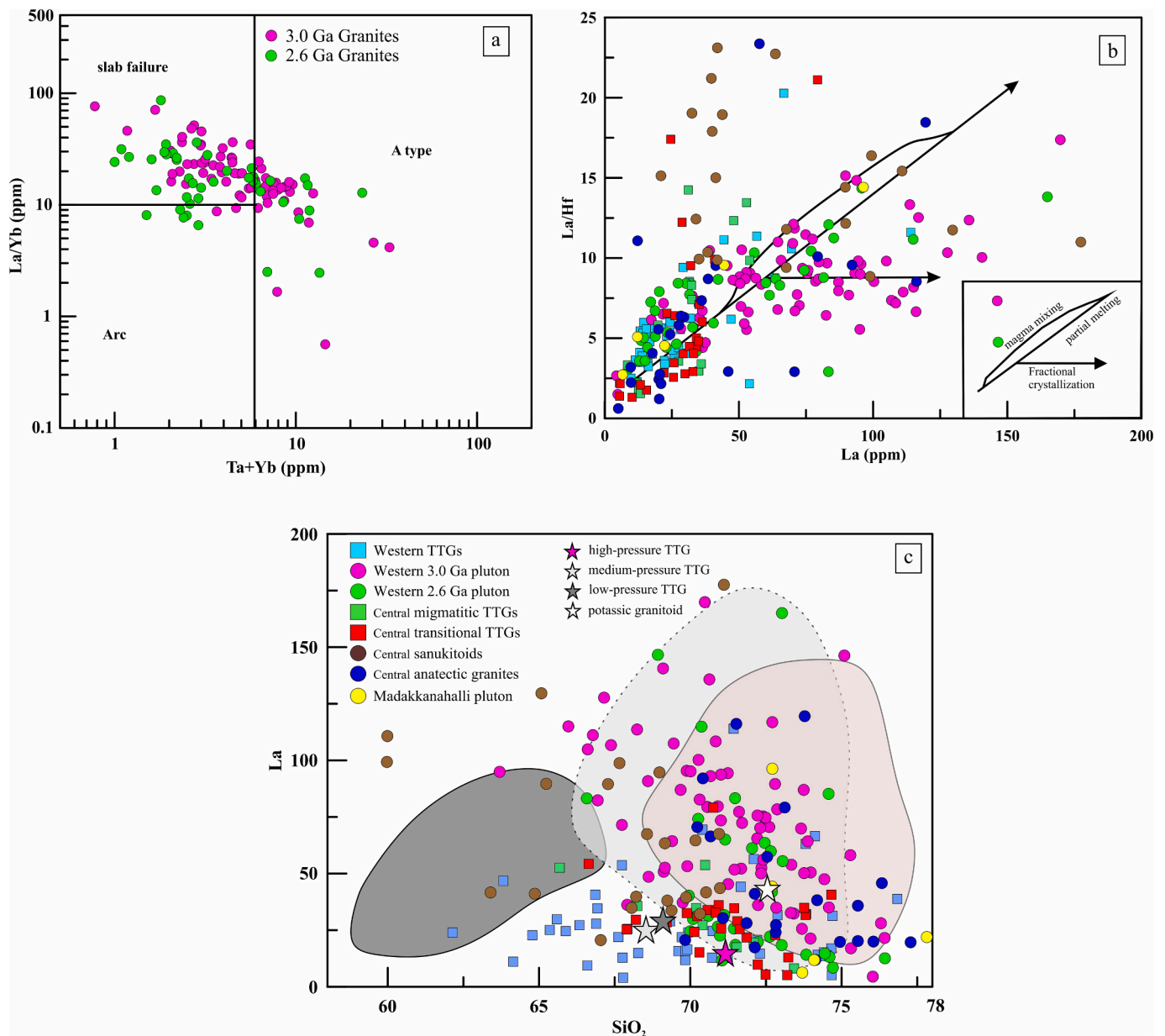


Fig. 32. (a) La/Yb versus Ta+Yb plot (after Whalen and Hilderbrand, 2019) for ca.3.0 and ca. 2.6 Ga plutons suggest slab failure associated with collision of arcs (b) La/Hf versus La plot (Abdallsamed et al., 2018) showing the studied granitoids from western and eastern basement follow the partial melting and magma mixing trend (c) La versus SiO₂ (after Simon et al., 2021) for western and eastern basement granitoids.

11.3. Eastern corridor

11.3.1. Javagondanahalli-Kunigal volcanics

Field evidence such as dominant mafic volcanics with pillow structures suggest their eruption in marine environments. The narrow range of SiO_2 (46–52 wt.%) combined with weak trends on the Harker's binary plots (Figure is not given) argue against fractional crystallization of their parental magmas. This is in conformity with La/Sm versus La plot (see Fig. 27b after Schiano et al., 2010) and Zr/Sm versus Nb/Sm (see Fig. 28c, Bai et al., 2022) where majority of the samples of the Javagondanahalli-Kunigal volcanics do not follow differentiation trends rather suggest melts originated from composite sources with crustal contamination. Elemental characteristics like higher alkalis (average $\text{Na}_2\text{O} = 2.17$; $\text{K}_2\text{O} = 0.33$) coupled with variable but generally higher incompatible element contents ($\Sigma\text{REE} = 28.43$ to 68.02 ppm; $\text{Rb} = 1.45$ – 13.46 ; $\text{Ba} = 9.6$ – 133.03 ; $\text{Sr} = 36.02$ – 243.21) indicate heterogeneous source reservoirs. Incompatible elemental ratios are used to characterize compositions of potential mantle source reservoirs. On La/Sm versus Sm/Yb plot (see Fig. 27c, Aldanmaz et al., 2000), the Javagondanahalli volcanics consistent with their derivation from highly heterogeneous mantle reservoirs ranging in composition from depleted to primitive mantle to E-MORB sources with few samples plotting towards N-MORB. Other incompatible element ratios like Th/Nb vs TiO_2/Yb (Fig. 29c, Pearce et al., 2021) and Sm/Yb vs Sm (see Fig. 27d, Aldanmaz et al., 2000) are also consistent with the origin of mafic volcanics from depleted mantle to primitive mantle reservoirs under spinel-lherzolite to garnet-lherzolite stability conditions. The origin of heterogeneous arc mantle reservoirs is not clear. It can either represent enrichment linked to upward flow of fluids liberated by slab dehydration or melting of sediment components in the subducting slab. However, elemental ratios such as Th/Nb vs Ba/Th (Fig. 29d, Saccani et al., 2018), and $(\text{Hf}/\text{Sm})_{\text{N}}$ vs $(\text{Nb}/\text{La})_{\text{N}}$ (see Fig. 29a, Hu et al., 2021) depict the involvement of both sediment-derived melts and subducting slab derived fluids in the enrichment of the mantle, which is in agreement with Ba/Th versus Th plot (see Fig. 29b, La Flèche et al., 1998). Neodymium isotope data of Javagondanahalli [$\epsilon_{\text{Nd(T)}} = +3.8$ to -0.4] also indicate the involvement of heterogeneous mantle sources (depleted to primitive mantle) with traces of ancient crustal contamination.

11.4. Origin of granitoids: Western corridor

11.4.1. ca. 3350–3200 Ma TTG-type gneisses

Numerous studies focused on the origin TTGs from the Archean cratons worldwide (e.g. Martin and Moyen, 2002; Guitreau et al., 2012; Laurent et al., 2014, 2020; Jayananda et al., 2018, 2023; Palin et al., 2016; Deng et al., 2019; Pourteau et al., 2020; Marimon et al., 2022 and references therein). Many studies have shown that Archean TTGs originated by hydrous melting of compositionally diverse mafic sources at different pressures with corresponding depths (Martin et al., 2014; Moyen, 2011; Moyen and Martin, 2012; Jayananda et al., 2018). The ca. 3330–3200 Ma old TTGs show dominant high-Al character with few samples with low-Al contents compared to the TTGs in the cratonic core (Jayananda et al., 2023). The wide range of SiO_2 (62.13–74.65 wt.%) coupled with moderate to strong negative trends of Al_2O_3 , TiO_2 , Fe_2O_3 , MgO , CaO and P_2O_5 on Harker's binary diagrams suggest the role of magmatic differentiation wherein plagioclase, hornblende, accessory titanite, ilmenite and apatite represent fractionating phases or residual phases in source. The high SiO_2 (>62.0 to 74.0 wt.%), Na_2O (>4.0 wt.%), low ferro-magnesian elements ($\text{FeO} + \text{MgO} < 4.0$ wt.%), K_2O (<2.0 wt.%) and higher incompatible element contents (Rb, Ba, Sr, Th, U, Y, Zr, Hf and REE) preclude their derivation by melting oceanic crust that originated at a spreading centre. Melting of oceanic crust produced at spreading centres cannot explain the observed high incompatible elements in the studied TTGs (Hoffmann et al., 2011; Martin et al., 2014; Jayananda et al., 2015, 2018). Melting of oceanic plateau similar to komatiite-to-komatiitic basalts preserved in the adjacent greenstone

belts (e.g. ca. 3384 Ma old J. C. Pura belt) can be considered. However, those komatiites are also highly depleted in incompatible elements and melts derived from such lithologies cannot account for the high incompatible element contents in the studied TTGs. Furthermore, melting of oceanic plateaus requires very high temperatures and melts produced would have to undergo large degrees of differentiation to produce a small volume of TTGs. No evidence of melting of komatiite and komatiitic basalts in the J. C. Pura, Ghattihsalli and Banasandra greenstone belts is observed. Consequently, oceanic crust similar to MORB composition or oceanic plateaus as potential sources is ruled out. Alternatively, melting of thickened oceanic arc crust produced in shallow angle subduction (e.g., Polat, 2012; Hoffmann et al., 2011; Nagel et al., 2012; Jayananda et al., 2015) can be considered. The variable but generally high incompatible element contents (Rb, Ba, Sr, REE, Nb, Ta, Zr, Hf, Y) of the studied TTGs are compatible with their derivation from heterogeneous source reservoirs. This inference is in conformity with source discriminate triangular plot $\text{Al}_2\text{O}_3/\text{FeO} + \text{MgO} - 3\text{CaO} - 5\text{K}_2\text{O}/\text{Na}_2\text{O}$ (Laurent et al., 2014) where the studied TTGs fall on low-K to high-K mafic source with minor involvement of pre-existing low-K tonalitic crust (Fig. 30a). The strong negative Ba-Sr, Nb-Ta, Ti anomalies on primitive mantle normalized spider diagrams suggest retention of plagioclase, garnet, amphibole and ilmenite in the source. Incompatible elements and their ratios like $(\text{La}/\text{Yb})_{\text{N}}$ versus $(\text{Yb})_{\text{N}}$ (Jahn et al., 1981; Fig. 30b) and Sr/Y versus Y (Defant and Drummond, 1990; Fig. 30c), reveal melting sources at different depths with variable residual garnet (7–30%) and/or amphibole. Pressure sensitive elements like Nb versus Y (after Simon et al., 2021, Fig. 30d), Ce/Sr versus Y (after Moyen, 2011, Fig. 31a) also suggest an origin of TTG magmas under medium to high pressure conditions which is in conformity with Sr/Y versus $\text{K}_2\text{O}/(\text{Na}_2\text{O} + \text{CaO})$ plot (Hoffmann et al., 2011, Fig. 31b) indicating that TTG magmas formed by melting of heterogeneous mafic source at moderate to high pressure (8–20 kb) conditions. The published Nd isotope data (Chardon, 1997; Sebastian et al., 2021) coupled with our new data ($\epsilon_{\text{Nd(T)}} = +2.3$ to -0.7) indicates a major juvenile source (depleted to chondritic reservoirs) with traces of pre-existing crust in the origin of TTGs (see Fig. 25a). This is consistent with Hf isotope data ($\epsilon_{\text{Hf(T)}} = 5.9$ to 3.5) which suggest derivation of TTG magmas from juvenile sources possibly newly formed thickened arc crust (see Fig. 25b). To summarize, TTGs forming the basement along the western margin of the Chitradurga greenstone belt are originated by moderate to high pressure melting of juvenile sources (depleted to enriched) with pre-existing crustal input.

11.4.2. ca. 3000–2950 Ma potassic granites

The ca. 3000–2950 Ma Bellur-Nagamangala, Hosdurga and Bukkapatna granite plutons intrude the Archean basement comprising TTGs and interlayered older Sargur Group greenstone. The origin of ca. 3000 Ma Bellur-Nagamangala granites in the western Dharwar craton is attributed to melting of pre-existing TTGs in the lower crustal levels (1.2 GPa) with minor juvenile mafic magmatic input (Jayananda et al., 2019). On the contrary, Sebastian et al. (2021) based on elemental and Nd isotope [$\epsilon_{\text{Nd(T)}} = -2.9$ to -7.6] and Sr [$I_{\text{Sr}} = 0.715$ – 0.725] isotope data interpreted the origin of ca. 3000 Ma Hosdurga granites as product of low degree (11%) melting of pre-existing TTGs. However, their interpretation of low degree melting of pre-existing TTG source based on calculation of $\epsilon_{\text{Nd(T)}}$ values at 2600 Ma. However, U-Pb zircon age in this study indicates 2955 ± 5 Ma which in turn change the $\epsilon_{\text{Nd(T)}}$ values which are more compatible with a hybrid source (see Fig. 25a).

The observed moderate to strong negative trends of TiO_2 , Al_2O_3 , Fe_2O_3 , MgO , CaO , P_2O_5 against SiO_2 suggest a possible role of plagioclase, hornblende or orthopyroxene, ilmenite and apatite as fractionating phases or interaction of felsic to intermediate magmas originated at different depths. Alternatively, such trends could be related to the retention of plagioclase, hornblende and/or orthopyroxene, ilmenite and apatite as residual phases during partial melting of source rocks.

The key petrological features of ca. 3000 Ma potassic granites

including co-existing felsic to intermediate magmas, large kyanite-muscovite xenoliths, magmatic epidote along with euhedral quartz-feldspar crystals suggest the rapid emplacement of magmas from different depths into the thickened crust. The major and trace element characteristics like high K_2O , Fe/Mg , Rb , Zr , Hf , Nb , Ta , Y , REE with low CaO show affinity to A-type granites. Various petrogenetic models proposed for the origin of A-type granites including (i) fractionation of mantle derived magmas to evolve into late-stage alkali-rich fraction (Loiselle and Wones, 1979); (ii) late-stage alkali, Cl, F rich magma interaction with host granites (Harris et al., 1982); and (iii) melting of high grade felsic granulitic source at lower crustal levels (Collins et al., 1982).

The major and trace element data of granites preclude their origin through differentiation of mafic magmas as such process needs huge volumes of mafic magmas to undergo fractionation and no contemporaneous mafic rocks or ultramafic cumulates were found. Further, observed petrographic and elemental data are not consistent with interactions and metasomatic alteration of host granites through fluid flow. Additionally, melting of lower crustal felsic granulitic source can generate small quantities of alkaline granite and cannot explain large volumes of the ca. 3000 Ma granite intrusions in the studied corridor. Their high potassic character ($K_2O/Na_2O > 1.1$ in most studied samples) coupled with lower CaO , moderate to very high abundances of incompatible elements (Rb , Ba , Th , U , Nb , Ta , Zr , Hf , Y , and REE) preclude their derivation from depleted sources. Alternatively, remelting of pre-existing TTG crust or melting of hybrid sources involving pre-existing TTG crust and enriched mafic rocks can be considered. Biotite-muscovite bearing granites originated from melting of TTG crust are characterized by lower [$Zr < 200$ ppm; $Hf < 5$ ppm; $Nb < 8$ ppm; $Ta < 0.52$ ppm and $Y < 15$ ppm; Laurent et al., 2014; Jayananda et al., 2006, 2020; Wang et al., 2020] whilst most of our studied samples display higher $Zr > 200$ –800 ppm; Hf 4–14 ppm; Nb 10–26 ppm (except few samples show less than 10 ppm); Ta 1–2.4 ppm (except few samples < 0.1 ppm) and Y 20–309 ppm (except few samples with less than 20 ppm). These characteristics preclude their origin by melting of pre-existing TTG crust alone. On the magma source discrimination triangular plot $2^*A/CNK-Na_2O/K_2O-2FMSB$ (Laurent et al., 2014) most samples plot in hybrid source involving ancient TTGs and enriched mafic rocks (see Fig. 20b). This is in consistent with source discrimination $Al_2O_3/FeO+MgO-3CaO-5K_2O/Na_2O$ plot (Laurent et al., 2014) where these potassic granites plot on TTG to high-K mafic source with traces of sedimentary input (see Fig. 30a). The high total REE (85.99–826.41 ppm) coupled with poorly to highly fractionated REE patterns coupled with negative Eu anomalies [$(Eu/Eu^*) = 0.26$ –0.67] suggest an origin of granite magmas from mixed sources (pre-existing TTGs and enriched mafic rocks) with retention of plagioclase in residue with. This observation is in agreement with observed strong negative Ba, Na-Ta, Sr, Ti anomalies and positive Hf, Zr and Y on the primitive mantle normalized multi-element diagrams preclude residual garnet whilst retention of plagioclase, titanite and rutile in the residue. Our whole-rock Nd isotope and in-situ Hf isotope data in zircons together with published record ($\epsilon_{Nd(T)} = +1.9$ to -3.9) and [$\epsilon_{Hf(T)} = -4.0$ to $+4.3$] is consistent with hybrid sources involving pre-existing TTGs and juvenile mafic rocks (Jayananda et al., 2019; Ao et al., 2021; Sebastian et al., 2021). Incompatible elemental ratios of potassic granites with A-type affinity have been attributed to arc - slab failure to anorogenic settings (Whalen et al., 1987; Pearce et al., 1984; Whalen and Hildebrand, 2019). Elemental ratios of studied ca. 3000 Ma potassic granites projected on the La/Yb versus Nb+Y (Fig. 31c after Whalen and Hildebrand, 2019) imply a geodynamic setting from slab failure (slab break off) probably associated with collision of arc with continent lead to development of a hotspot which is in conformity with Rb versus Nb+Y (Fig. 31d after Pearce et al., 1984) and La/Yb versus Ta+Yb plot (Fig. 32a after Whalen and Hildebrand, 2019).

11.4.3. ca.2600 Ma potassic granites

The 2600 Ma plutons include elongated Chitradurga granite pluton in the north and Pandavapura granite pluton in the south intruding Chitradurga greenstone belt and Arsikere-Banavara granite plutons show a wide range of elemental and Nd-Sr isotopic compositions. Their origin is attributed to melting of pre-existing crustal rocks at different depths (Jayananda et al., 2006; Mohan et al., 2023; Sebastian et al., 2023). These granites show slightly higher contents of TiO_2 , Fe_2O_3 , MgO , P_2O_5 , Nb , Y and lower Al_2O_3 , Na_2O , Rb , Sr , Zr for the similar SiO_2 contents compared to ca. 3000–2950 Ma potassic granites. The moderate to strong negative trends of Al_2O_3 , TiO_2 , Fe_2O_3 , MgO , CaO , P_2O_5 , Sr , Y and V reveal the possible role of plagioclase, hornblende, orthopyroxene, clinopyroxene, ilmenite and apatite as fractionating phases or residual phases in the source. The La/Hf versus La (after Abdallsamed et al., 2018, Fig. 32b) used to evaluate the role of fractional crystallization, magma mixing and partial melting to explain the observed chemical variation on which the studied granites follow the partial melting and magma mixing trend. The large spread on La versus SiO_2 plot (Fig. 32c after Simon et al., 2021) preclude fractional crystallization as a major process for observed chemical variation. This is consistent with poorly to moderately fractionated REE patterns [$(La/Yb)_N = 1.65$ –23.92], negative Eu anomalies ($Eu/Eu^* = 0.15$ –0.65) flat HREE coupled with strong negative anomalies of Ba, Nb, Ta, Sr, Ti on primitive mantle normalized multi-element diagram (see Fig. 23b) suggest retention of plagioclase, hornblende, ilmenite with minor or without garnet in source. The wide range of SiO_2 (66.7–76.4 wt.%), Al_2O_3 (10.8–17.2 wt.%), MgO (0.05–1.8 wt.%), alkalis (5.8–11.4 wt.%), REE (49.8 to 749.6 ppm) coupled with their potassic character extend into transitional TTGs on $2^*A/CNK-Na_2O/K_2O-2FMSB$ (after Laurent et al., 2014; see Fig. 20b) which suggest dominant crustal source with minor enriched juvenile mafic crust input. This is in conformity with triangular plot of $Al_2O_3/FeO+MgO-3CaO-5K_2O/Na_2O$ (after Laurent et al., 2014; see Fig. 30a) wherein the ca. 2600 Ma granites plot in high-K mafic rocks to low-K tonalite with few samples extending to metasedimentary source. Therefore, it appears that these granites originated from melting of dominant lower crustal TTGs with minor enriched juvenile mafic crust similar Chitradurga mafic volcanics. Nd isotope data [$\epsilon_{Nd(T)} = -1.3$ to -6.2 excluding anomalous values] also indicate their derivation from sources involving dominantly pre-existing crustal source with minor juvenile input (Jayananda et al., 2006; Sebastian et al., 2023; Mohan et al., 2023). Elemental and Nd isotope data reveal sources involving major crustal input in their genesis, but the depth of melting is not clear. Pressure sensitive elements are used to evaluate the depth of melting (Hoffmann et al., 2011; Moyen, 2011; Laurent et al., 2014). On Sr/Y versus (K_2O/Na_2O+CaO) plot (Hoffmann et al., 2011) most of the samples indicate 7–15 Kbar corresponding to melting in the middle crust to thickened lower crust (see Fig. 31b).

The tectonic context of their origin and emplacement is not clear. The Chitradurga pluton intrudes the greenstone sequences and adjacent TTGs and ca. 2965 Ma granitoid gneiss southeast of Chitradurga town (Mohan et al., 2023). Our new field data reveal a continuity of the ca. 2740 Ma Javagondanahalli greenstone belt (see Fig. 1b) towards the north beyond Chitradurga. In this scenario, Chitradurga granite found between two volcanic arcs is probably related to a collision of Chitradurga and Javagondanahalli arc close to 2620 Ma. East of Chitradurga, Lu-Hf is-situ dating of garnet from high pressure assemblages (14–13 Kbar; Kaempfer et al., 2024) of retrograded eclogite along NW trending shear zone reveal probably corresponds to tectonically transported oceanic lithosphere during arc collision. The tectonic framework of ca. 2620 Ma Arsikere - Banavara granites (see Fig. 1b) intruding old (> 3100 Ma) basement imply anorogenic setting linked to mantle upwelling/hot spot driven by slab breakoff associated with westward subducting crust beneath TTG basement.

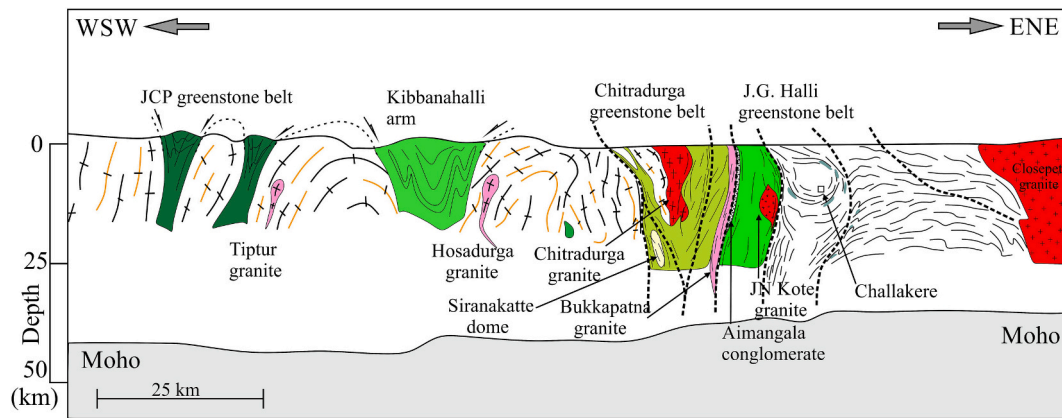


Fig. 33. Interpretative cross-section of WSW-ENE crustal corridor across the Chitradurga boundary shear showing the crustal architecture including major tectonic units, fabrics and lithological assemblages.

11.5. Origin of granitoids: Eastern corridor

11.5.1. ca. 3338-3176 Ma migmatitic TTGs

The TTGs show similar elemental and whole-rock Nd isotopic characteristics compared to the TTGs in the western corridor as reflected in several major and trace element plots (see Figs. 25a) and also define moderate to strong differentiating trends. Therefore, these migmatitic gneisses possibly originated by similar petrogenetic mechanisms that involve differentiation of parental magma which originated from moderate to high pressure melting of mafic arc crust with retention of variable plagioclase, hornblende, ilmenite with or without garnet in the source residue (Jayananda et al., 2015, 2023). These rocks show similarities in major and trace element characteristics compared to western block TTGs in particular low- Al_2O_3 and moderate to poorly fractionated REE patterns $[\text{La}/\text{Yb}]_{\text{N}} < 10$ which reflects their origin by low to moderate pressure melting of mafic arc crust with retention of plagioclase, hornblende, ilmenite but without significant garnet in residue.

11.5.2. ca. 2670-2600 Ma transitional TTGs

Published ages (ca. 2630-2700 Ma) reveal that transitional TTG accretion was either contemporaneous or slightly post-date (10-50 Ma) the ca. 2676-2700 Ma greenstone volcanism in the adjacent Ramagiri-Penakacherla and Sandur greenstone belts in the studied corridor (Nutmán et al., 1996; Balakrishnan et al., 1999; Naqvi et al., 2002). Elemental data (Dey et al., 2016; Jayananda et al., 2020; this study) reveal a wide range of SiO_2 (55-70 wt.%), higher CaO, K_2O , MgO coupled with higher contents of both compatible (Ni, Cr) and incompatible elements (Rb, Ba, Sr, Nb, Ta, Zr, Hf, REE) compared to >3200 Ma TTGs from the western Dharwar craton. These characteristics coupled with their transition from trondhjemite to calc-alkaline differentiation trend (see Fig. 20a-c) argue for their origin by distinct petrogenetic processes. Moderate to strong negative correlation of TiO_2 , Al_2O_3 , Fe_2O_3 , MgO and CaO on Harker's binary plots suggest variable role of plagioclase, hornblende, ilmenite as fractionating or residual phases. The high CaO, MgO, K_2O low to high total REE combined with Nd isotope data cannot be explained either by melting of oceanic crust produced at ridges or by arc crust or >3000 Ma ancient crust alone. Alternatively composite sources involving oceanic arc crust and heterogeneous mantle (enriched to depleted). The observed low to high total REE (29.31-493.02 ppm), moderate to highly fractionated REE patterns $[(\text{La}/\text{Yb})_{\text{N}} = 14.16 - 57.68]$; present study; Jayananda et al., 2020) with minor or without significant Eu anomalies $(\text{Eu}/\text{Eu}^* = 1.42-0.74)$ reveal their derivation from composite sources with variable residual plagioclase, hornblende, garnet and ilmenite. The Nd isotope data $[\epsilon_{\text{Nd(T)}} = +2.0 \text{ to } -1.6]$ including Nd model ages of 3160 – 2700 Ma, older than crystallization ages, suggest their origin from source reservoirs involving pre-existing crust (arc crust and minor old crust) and heterogeneous

mantle. The enrichment of incompatible elements in the sources of transitional TTGs is attributed to the melting of subducted sediments derived from rapid chemical weathering intervening formed Neoproterozoic volcanic arcs and ascent of those melts cause eventual enrichment in the overlying arc mantle. Melting of such composite sources involving arc mantle and newly formed arc crust, with or without traces of old crust, can generate magmatic precursors of the transitional TTGs. This is in agreement with observations made by Jayananda et al. (2020) who argued that transitional TTGs in the Eastern Dharwar craton originated from melting of composite sources that involved arc crust, and mantle at different depths. The higher K_2O , Ba, Sr, coupled with higher HFS including REE, MgO, Ni, Cr coupled Nd isotope $[\epsilon_{\text{Nd(T)}} = +2.0 \text{ to } -1.6]$ and Hf isotopes $[\epsilon_{\text{Hf(T)}} = +6.5 \text{ to } -3.0]$; present study; Jayananda et al., 2020 compared to the ca. 3200 Ma TTGs also points to composite sources involving arc crust, enriched mantle, and with minor or without ancient crustal input.

11.5.3. ca. 2560-2530 Ma sanukitoids

Sanukitoids are widespread and voluminous calc-alkaline rocks occur as large north-south trending Closepet batholith which contribute to major Neoproterozoic crustal growth. They are extensively studied by several workers (Dey et al., 2016; Jayananda et al., 1995, 2000, 2020; Moyen et al., 2001, 2003; Sarvothaman, 2001). On the $2^*\text{A}/\text{CNK}-\text{Na}_2\text{O}/\text{K}_2\text{O}-2\text{FMSB}$ triangular plot (Laurent et al., 2014) for classification of late Archean granitoids all the studied rocks plot in the sanukitoid field. Their wide range of SiO_2 (52-70 wt.%) and moderate to strong negative trends of TiO_2 , Al_2O_3 , Fe_2O_3 , MgO and CaO on Harker's binary plots are attributed to fractionation of plagioclase, hornblende/clinopyroxene, ilmenite (Jayananda et al., 2018, 2020; Figure not given). Their moderate to strongly fractionated REE patterns $[(\text{La}/\text{Yb})_{\text{N}} = 10.84-47.23]$ with mild negative or without Eu anomalies $[\text{Eu}/\text{Eu}^* = 0.7490-0.98]$ suggest either simultaneous crystallization of plagioclase and hornblende or retention of variable plagioclase, hornblende and/or garnet in source residue. On the other hand, negative Nb-Ta-Ti anomalies on the primitive mantle normalized multi-element diagrams indicate retention of ilmenite/titanite whilst positive to negative Ba-Sr-Y anomalies suggest variable (with or without) residual plagioclase, hornblende and/or garnet in the source. The high MgO, Ni, Cr cannot be explained by remelting of pre-existing crustal source as crustal melts are characterized by much lower MgO (<0.20 wt.%), Ni, Cr but conversely high SiO_2 (>70 wt.%). On the $\text{Al}_2\text{O}_3/\text{FeO}+\text{MgO}-3\text{CaO}-5\text{K}_2\text{O}/\text{Na}_2\text{O}$ triangular plot (Laurent et al., 2014) the studied sanukitoids plot in the enriched mafic source field (see Fig. 30a) which is compatible with high contents of both incompatible (K_2O , Rb, Ba, Sr, Zr, Hf, REE, Y) and compatible (MgO, Ni, Cr) elements (supplementary table 4) and also reflected on $1/\text{Er}-\text{Ba}+\text{Sr}/1000-\text{Er}$ plot (figure is not given) where sanukitoids plot in the enriched mantle field (Heilimo et al., 2010). Such source is

consistent with the whole-rock Nd isotope [$(\epsilon_{\text{Nd}(T)} = +2.0$ to 0.0 with few negative values up to $-2.6)$] and in-situ Hf data [$(\epsilon_{\text{Hf}(T)} = +4.5$ to -3.2 ; Jayananda et al., 2020) which indicate their origin from enriched mantle source reservoirs with minor crustal input (Jayananda et al., 2018). Nd isotopic and elemental data are consistent with enrichment of the mantle through slab- and/or sediment-derived melts in subduction zones (Jayananda et al., 2020). Consequently, sanukitoids were generated by melting of enriched mantle reservoirs with minor crustal input forming a major juvenile magmatic addition that contributed to terminal Archean crustal growth during 2560–2520 Ma.

11.5.4. ca. 2507 Ma Maddakkanahalli granite

The aplitic ca. 2500 Ma Maddakkanahalli granite (Chardon et al., 2011) is the youngest Archean magmatic intrusion in the Dharwar craton found at the contact zone of Javagondanahalli greenstone belt and basement TTGs. It is found as large sheets along flat foliations of the transitional TTGs and veins in the Javagondanahalli greenstone belt. This aplitic granite contains abundant quartz and alkali feldspar with traces of biotite, muscovite, rare zircon and titanite. The high content of SiO_2 (72.69–76.80 wt.%) K_2O (4.1–6.48 wt.%), very low content of ferromagnesian elements together with moderate total REE, poorly fractionated REE patterns coupled with negative Eu anomalies point to its possible origin by melting of pre-existing crust. This is consistent with its whole-rock Nd isotope data [$(\epsilon_{\text{Nd}(T=2500 \text{ Ma})} = -2.7$ to $-6.8)$]. In summary, Maddakkanahalli granite formed by melting of pre-existing crust and intruded into the flat to shallow fabrics of the transitional TTGs, migmatitic gneisses, and Javagondanahalli greenstone belt.

12. Discussion

12.1. Distinct crustal architecture and magmatic records across the Chitradurga boundary shear zone

The crustal architecture, lithologic assemblages, metamorphic and magmatic records across the boundary shear zone particularly the existence of two crustal blocks with distinct crustal histories and polarity of Neoproterozoic subduction have been subjects of much discussion in recent years (Dey et al., 2016; Bidyananda et al., 2016; Jayananda et al., 2013a, 2013b, 2018, 2020; Krapez et al., 2020; Mandal et al., 2018; Vijaya Rao et al., 2015; Mohan et al., 2014, 2020, 2023). The studied crustal corridor shows a fundamental difference in crustal architecture across the Chitradurga boundary shear zone (Chardon et al., 1996, 2008, 2011). West of Chitradurga greenstone belt, the crust is thick (42–51 km; Gupta et al., 2003) and tectonic fabric data and kinematic analysis reveal spectacular dome-basin patterns which are bordered by widely spaced regional north trending shear zones (Bouhallier et al., 1993, 1995; Chardon et al., 1996). In contrast, immediately to the east of the Chitradurga boundary shear zone, spectacular flat to shallow foliations linked to lower crustal flow (without accommodating thickening) are associated with wide-spread hot juvenile magmatic additions into the lower crust during ca. 2650–2540 Ma, causing lateral flow preventing crustal thickening (Chardon et al., 2011; Jayananda et al., 2013a, 2020) which is consistent with thinner crust of 32–35 km (Fig. 33).

The corridor west of the Chitradurga greenstone belt preserves 3500–3200 Ma TTG-komatiite dominated greenstone assemblages forming the basement which in turn was intruded by ca. 3000 Ma and ca. 2600 Ma high potassic granitoids (Guitreau et al., 2017; Jayananda et al., 2006, 2008, 2015, 2019, 2023; Ranjan et al., 2020; Ao et al., 2021; Maya et al., 2017; Mukherjee et al., 2012; Wang et al., 2023; Mohan et al., 2014, 2023; Ravindran et al., 2023). The ca. 3000–2850 Ma intracratonic basins developed on the eroded basement with paleosol at their contacts with the basement. On the contrary, the eastern corridor contains large remnants of ca. 3300–3176 Ma migmatitic TTGs with widespread ca. 2690–2600 Ma transitional TTGs associated with bimodal volcanics (ca. 2700 Ma boninite-adakite-basalt-felsic volcanics and 2540 Ma pyroclastic felsic flows) that was intruded by voluminous ca. 2560–2530 Ma

sanukitoids (Manikyamba et al., 2005, 2012, 2015; Naqvi et al., 2002; Bidyananda et al., 2016; Khanna et al., 2016; Dey et al., 2016; Jayananda et al., 2000; Jayananda et al., 2008; Jayananda et al., 2020). Although ca. 2700–2600 Ma volcano-sedimentary greenstone sequences are documented both in western and eastern corridors, fundamental differences do exist in the nature and composition of volcano-sedimentary assemblages (Jayananda et al., 2013a). The western corridor contains basalt-pyroclastic flows with voluminous shallow to deep water sediments whilst eastern corridor displays dominant volcanics with bimodal age distribution containing basalt-boninite-adakite-rhyolite assemblages with minor sediments (Manikyamba et al., 2012, 2015).

Reworking of pre-existing crust leading to the generation of potassic magmas and their emplacement caused the stabilization of Archean continental crust in the western corridor during ca. 3000–2950 Ma and 2600 Ma (Jayananda et al., 2019) whilst such crustal reworking events and potassic magmatism is not known in the Eastern corridor. However, widespread juvenile magmatism during ca. 2560–2520 Ma (Jayananda et al., 2020) led to the emplacement of calc-alkaline batholiths (e.g., Closepet batholith).

12.2. Contrasting metamorphic records in crustal domains across boundary shear zone

The thermal history of the Archean crust across the Chitradurga boundary shear zone is not well understood as not many studies focused on metamorphic history. Raase et al. (1986) documented a progressive north to south increase of pressure - temperature along Chitradurga belt from greenschist to hornblende-granulite facies whilst Pichamuthu (1965) documented an increase of metamorphic pressure -temperature from north to south along the Closepet granite. On the contrary, Rollinson et al. (1981) described contrasting metamorphic assemblages in the southern parts of the western and eastern blocks of the Dharwar craton. Subsequent studies have shown contrasting thermal records of the two cratonic blocks across the boundary shear zone (Jayananda et al., 2008; Jayananda et al., 2011; Jayananda et al., 2013b; Kaempf et al., 2024). Along the western boundary of Chitradurga greenstone belt, discontinuous outcrops of kyanite-muscovite bearing assemblages occur in the gneisses in the Kudineerkatte in the north and Nelligere in the south (see Fig. 2c). Immediately west, in the Kalyadi region, ca. 2975 Ma old garnet-kyanite bearing metapelite indicates pressure-temperature conditions of 10.9–11.3 kb and 685–700°C. On the contrary, along the eastern boundary of Javagondanahalli greenstone belt, cordierite-orthopyroxene-anthophyllite bearing assemblages indicate low pressure and higher temperature (4.0 - 4.5 kb and 700–750°C; Jayananda et al., 2013a) and further east in the metapelite from eastern boundary of Closepet batholith reveal 6.7–5.0 kb and 930 - 820°C (Jayananda et al., 2011; Kaempf et al., 2024).

Adjacent to the Chitradurga boundary shear zone (about 30 km ENE of Chitradurga), ca. 2630 Ma high pressure assemblages represent tectonically transported fragments probably linked to the collision of arcs and emplacement of Chitradurga-Jampalnakankote granites. The ca. 2630 Ma retrograded eclogite (garnet-clinopyroxene-actinolite) is associated with a NW trending sinistral shear zone to the east of Javagondanahalli greenstone belt. This metamorphic age also coincides with the age of associated host transitional TTGs. Furthermore, sodic mica bearing pelitic rock (19CH01) in this study was collected from NNW trending shear zone traversing the central part of the Chitradurga belt (see Fig. 1b). The sodic mica bearing high pressure (~13 kb) pelitic assemblage (see Fig. 1b) probably represents a reactivated thrust developed during collision of two volcanic arcs of the Chitradurga belt.

Therefore, we speculate that the retrograded eclogites and sodic mica bearing assemblages could represent material tectonically transported through back thrusting during the collision of transitional TTGs magmatic arc with Javagondanahalli volcanic arc and volcanic arcs of Chitradurga belt through WSW convergence of intervening oceanic

lithosphere.

12.3. Successive stages of continental growth and mantle evolution: Insights from granitoids and greenstone volcanics

12.3.1. Constraints from zircon U-Pb ages, whole-rock Nd and in-situ Hf isotopes

The preserved Archean continental crust across the Chitradurga boundary shear zone formed through successive stages of magmatism contributing to crustal growth. U-Pb zircon ages combined with whole-rock Nd isotopes and in-situ Hf isotope data in zircons coupled with incompatible element data of granitoids and greenstone volcanics suggest that the Archean continental crust formed during four major episodes ca. 3400–3270 Ma, ca. 3230–3150 Ma, ca. 2700–2600 Ma and 2560–2530 Ma (Jayananda et al., 2008, 2016, 2018, 2020; Lancaster et al., 2015; Ranjan et al., 2020; Bidyananda et al., 2016; Mohan et al., 2014, 2023; Ravindran et al., 2023; Sebastian et al., 2021; Manikyamba et al., 2021; Wang and Santosh, 2019). Remnants of still older continental crust formed during ca. 3600–3500 Ma documented further southwest in the Gorur-Holenarsipur region forming cratonic core (Guitreau et al., 2017; Jayananda et al., 2023).

West of the Chitradurga boundary shear zone (see Fig. 1b) the U-Pb zircon ages reveal voluminous TTG type granitoids emplaced during ca. 3350–3250 Ma which in turn were intruded by ca. 3200–3150 Ma diapiric trondhjemite (this study; Chardon, 1997; Ao et al., 2021; Jayananda et al., 2019; Mohan et al., 2023; Ravindran et al., 2023; Sebastian et al., 2021) contributing to major continental growth. Whole-rock Nd isotope data of TTG type granitoids [$\epsilon_{\text{Nd(T)}} = +2.9$ to -0.9] reveal their derivation from depleted mantle reservoirs with minor crustal contamination. Existence of such depleted mantle reservoirs imply extraction of incompatible elements from mantle during an earlier episode of differentiation at ca. 3400–3600 Ma. In-situ Hf isotope data [$\epsilon_{\text{Hf(T)}} = +5$ to -4 excluding anomalous values] of TTGs also consistent with their origin from depleted mantle reservoirs with minor ancient crustal input imply an earlier episode of crustal growth probably at ca. 3400–3600 Ma. This episode of continental growth coincides with crustal growth documented further west in the large part of western Dharwar craton (Bidyananda et al., 2016; Ravindran et al., 2021, 2024; Jayananda et al., 2015, 2023; Guitreau et al., 2017; Ranjan et al., 2020). These felsic crust forming events are spatially and temporally associated with widespread ca. 3400–3150 Ma komatiite-dominated greenstone volcanism as revealed by whole-rock Sm-Nd isochrons in the adjacent Sargur Group greenstone belts in Ghattihsahalli, J.C. Pura, Nuggihalli, Banasandra, Nagamangala greenstone belts (Jayananda et al., 2008, 2023; Tushipokla and Jayananda, 2013; Maya et al., 2017; Mukherjee et al., 2012). Incompatible element ratios such as Nb/Th versus Nb/U (see Fig. 26a) coupled with Nd isotope data of komatiites [$\epsilon_{\text{Nd(T)}} = +5.5$ to 1.0] reveal their origin from depleted deep mantle reservoirs which in turn imply an earlier episode of differentiation of deep mantle contributing to crustal growth. Consequently, isotopic age data show two major periods of differentiation of shallow to deep mantle reservoirs at ca. 3350–3230 Ma and ca. 3200–3150 Ma, thereby contributing to major crustal growth with possible earlier episode at ca. 3600–3400 Ma documented in cratonic core (Guitreau et al., 2017; Jayananda et al., 2023; Ravindran et al., 2023). Recent ^{142}Nd isotope record on the ca. 3600 Ma crustal materials in the Western Dharwar craton reveal distinct Hadean mantle differentiation events (Ravindran et al., 2024). Further, incompatible elemental ratios and isotope (whole-rock Nd and in-situ Hf in zircon) data of the granitoids and greenstone volcanics in the western block argue for widespread crustal growth and formation of complementary shallow and deep depleted mantle reservoirs. These Paleo-Mesoarchean crust forming events were immediately followed by crustal reworking, metamorphism, emplacement of potassic plutons and stabilization of the crust during ca. 3150 and 3000 Ma (present study; Jayananda et al., 2019; Ao et al., 2021; Wang et al., 2023; Kaempfer et al., 2025; Mohan et al., 2023) and ca. 2600 Ma (Chadwick et al., 2007;

Jayananda et al., 2006, 2019; Mohan et al., 2014, 2023).

In summary, published Nd isotope record reveals ca. 3400–3200 Ma widespread komatiite volcanism to the west of the boundary shear zone with contemporaneous TTG-type granitoids accretion (Jayananda et al., 2008, 2023; Mukherjee et al., 2012; Maya et al., 2017). This komatiite volcanism has significantly contributed to crustal growth. Nd isotope data ($\epsilon_{\text{Nd(T)}} = +5$ to $+1$ excluding anomalous values) coupled with incompatible elemental data of komatiites and high-Mg basalts suggest their origin from depleted deep mantle reservoirs with long history of depletion whilst associated TTGs originated from shallow depleted mantle reservoirs with long term depletion history. Such long-term depletion of both deep and shallow mantle reservoirs implies an earlier episode of mantle differentiation possibly ca. 3600–3700 Ma.

The ca. 2900–2850 Ma Bababudan Group greenstone volcanism in the Kibbanahalli arm (Venkata Dasu et al., 1991; Chardon et al., 1996), along the western margin of the Chitradurga belt (Krapez et al., 2020) and Yediyur belt where basaltic volcanics erupted on the stabilized granitoid basement contribute to significant crustal growth. Incompatible element ratios (see Fig. 26a–b) coupled with Nd isotope data [$\epsilon_{\text{Nd(T)}} = -0.2$ to -1.5 ; see Fig. 25a] indicate their extraction from primitive to depleted mantle with minor crustal contamination. The voluminous younger Chitradurga Group bimodal mafic to felsic volcanism in the back-arc setting developed in major pulses during 2740–2677 Ma (Kumar et al., 1996; Jayananda et al., 2013a; Hokada et al., 2013) represents significant addition of juvenile crust. Incompatible element ratios (see Fig. 26a–b) coupled with Nd isotope data [$\epsilon_{\text{Nd(T)}} = +3.9$ to $+1.4$ with few negative values ranging -1.1 to -4.1] reveal their origin from depleted to enriched mantle with crustal contamination. Existence of shallow depleted to enriched mantle reservoirs indicate earlier extraction of magmas causing depletion of mantle while enrichment of arc mantle possibly through slab fluids related during subduction (see Fig. 25a).

Immediately east of the Chitradurga boundary shear zone, the Javagondanahalli-Kunigal greenstone volcanism is contemporaneous with Chitradurga greenstone as revealed by Sm-Nd whole rock isochron age of 2740 Ma (present study). Incompatible element ratios (see Fig. 26a–b) coupled with whole-rock Nd isotope [$\epsilon_{\text{Nd(T)}} = +0.4$ to 0.0 ; with two anomalous negative values (-0.1 to -2.3) probably linked to contamination see Fig. 25a] reveals their origin from a slightly depleted to primitive mantle with traces of crustal contamination. Greenstone volcanism in the Chitradurga and Javagondanahalli belts contributed to a Neoarchean crustal growth with evolution of complementary mantle reservoirs through extraction of magmas from mantle (see Fig. 25a). The enriched signatures observed in some samples could possibly be linked to fertilization of the mantle through slab-derived fluids and/or sediment-derived melts (see Fig. 29a).

Zircon U-Pb data indicate remnants of TTG-type migmatitic TTGs formed during ca. 3400–3176 Ma (Chardon et al., 2011; Nasheeth et al., 2015; Jayananda et al., 2000, 2013b, 2020; Bidyananda et al., 2016; present study). Their incompatible elemental data coupled with Nd isotope data [$\epsilon_{\text{Nd(T)}} = +2.9$ to -0.14] suggest their origin from depleted mantle with traces of pre-existing crustal input. Published records coupled with our new U-Pb zircon ages of transitional TTGs show a widespread juvenile magmatic accretion during ca. 2700 – 2600 Ma (Jayananda et al., 2006, 2020; Mohan et al., 2023; Nasheeth et al., 2015; this study) which is contemporaneous or preceded by basalt – boninite – andesite – felsic volcanism representing widespread continental growth throughout the Eastern Dharwar Craton. Their elemental and whole-rock Nd isotopes [$\epsilon_{\text{Nd(T)}} = +4.0$ to $+0.50$] and in-situ Hf data [$\epsilon_{\text{Hf(T)}} = +5.7$ to $+1.0$ with few negative $\epsilon_{\text{Nd(T)}} = -3.0$; see Fig. 25a] of transitional TTGs are consistent with depleted to primitive mantle reservoirs with minor crustal input thus contributing to widespread continental growth. The whole-rock Nd isotopes together with in-situ Hf isotope data and incompatible element contents suggest enrichment of their mantle reservoirs through sediment input from pre-existing crust or newly formed volcanic arcs into subduction (Jayananda et al., 2020).

U-Pb zircon ages (2560–2520 Ma) together with whole-rock Nd isotope [$\epsilon_{\text{Nd}(T)} = +3.0$ to -2.6] and in-situ zircon Hf isotope [$\epsilon_{\text{Hf}(T)} = +4.5$ to -3.0] of widespread granitoids of a sanukitoid affinity reveal major Neoarchean continental growth (Jayananda et al., 1995, 2000, 2018, 2020; Dey et al., 2016; Mohan et al., 2013; Chardon et al., 2011). Incompatible element contents together with whole-rock Nd isotope data [$\epsilon_{\text{Nd}(T)} = +3.0$ to -2.6] and in-situ Hf [$\epsilon_{\text{Hf}(T)} = +2.9$ to -1.6] suggest dominant primitive to enriched mantle with minor depleted mantle. The involvement of minor depleted mantle in sanukitoids genesis reflect an earlier episode mantle differentiation during ca 2700–2650 Ma contributing to crustal growth through greenstone volcanism and transitional TTGs formation in arc settings. The enriched mantle reservoirs evolved through fluid related or melt related enrichment during the Neoarchean (ca. 2700–2600 Ma) in arc environments. During the Neoarchean, major crustal growth has contributed to the establishment of larger continental landmass above the sea level. The detritus derived through weathering of continents shifted to intervening subduction zones where either dehydration of down going sediment laden slab and/or slab melting with sediments result in fluid flux and/or enriched melts which upon their upward migration would enrich the overlying mantle. Melting of such source reservoirs can explain the enriched elemental characteristics of sanukitoids.

12.3.2. Constraints from detrital zircon U-Pb ages

Published U-Pb ages of detrital zircons from Archean sedimentary sequences provide precise constraints on timing of their provenance formation which in turn reveal successive stages of crust formation, continental growth and supercontinental cycles. Detrital zircons from sedimentary units in Sargur Group and Dharwar Supergroup including Bababudan and Chitradurga Groups across the boundary shear zone reveal prominent age grouping at ca. 3400–3300 Ma, 3250–3140 Ma, 3000–2950 Ma, 2700–2620 Ma, and 2587–2548 Ma (Hokada et al., 2013; Lancaster et al., 2015; Krapez et al., 2020; Khelen et al., 2020; Nasheeth et al., 2016; Ravindran et al., 2023; Wang and Santosh, 2019). West of the boundary shear zone, detrital zircons from Sargur Group Ghattihsahalli schist belt yielded ages of 3451 ± 16 Ma and 3397 ± 12 Ma (Ravindran et al., 2023) whilst ~150 km directly south in Nagamangala-Karighatta, detrital zircons from muscovite schist and fuchsite quartzite yielded zircon U-Pb Concordia upper-intercept ages of $3343 \pm$ Ma and 3266 ± 14 Ma, 3247 ± 11 Ma (Wang and Santosh, 2019). These ages coincide with U-Pb zircon ages of adjacent TTG-type granitoids (present study; Mohan et al., 2023; Ravindran et al., 2023; Jayananda et al., 2019) forming potential provenance. Monazites from a muscovite-kyanite quartz schist belonging to the Sargur Group along the western margin of the Chitradurga belt indicate an U-Pb age of 3082 Ma (Hokada et al., 2013) which is probably related to the closure of Sargur Group basin that in turn coincides with the documented ages of metamorphism in the cratonic core J.C. Pura and Holenarsipur region (Jayananda et al., 2023; Dasgupta et al., 2019). Furthermore, zircon U-Pb ages of potassic granites intruding the basement (Sargur Group and TTGs) along the western margin of Chitradurga belt indicate ages possibly related to continental collision associated with closure of the Sargur Group basins close to ca. 3000 Ma (Jayananda et al., 2019; present study) followed by post-collisional extension that led to opening of Bababudan Group basins (e.g. Kibbanahalli arm and western margin of the Chitradurga belt). Detrital monazite age ca. 2700 Ma (Hokada et al., 2013) corresponds to a younger event probably related to the closure of Bababudan Group basin. Observed detrital monazite U-Pb ages of 3082 Ma are probably linked to the final closure of Sargur Group basins, inverse diapirism, metamorphism and cratonic core formation further southwest in Gorur-Hassan-Holenarsipur region (Dasgupta et al., 2019; Jayananda et al., 2023). The ca. 3000–2950 Ma granitoids (Jayananda et al., 2019; Ravindran et al., 2023; Ao et al., 2021; present study) coincide with a high-pressure thermal event (Kaempfer et al., 2025) probably driven by a possible continental collision and post-collisional thermal relaxation in the western Dharwar craton that resulted in the opening of Bababudan

Group basins.

The Neralakatte conglomerate form the basal unit of the preserved Bababudan Group along the western margin of Chitradurga belt provided detrital U-Pb zircon ages of 3451–3140 Ma for the timing of potential source regions corresponding to emplacement of adjacent basement granitoids (Hokada et al., 2013; Lancaster et al., 2015; Krapez et al., 2020). The opening of the stratigraphically equivalent to Bababudan basin along the western margin of Chitradurga belt probably occurred in a rift environment on an eroded basement following the emplacement of ca. 3000 Ma Hosdurga granite pluton and deposition ceased at least by 2719 ± 7 Ma (Krapez et al., 2020) marking the closure of basin. This closure of basin is reflected further west in the opening of a new rift system causing the emplacement of ultramafic magmas and deposition of Kaldurga conglomerate initiated close 2703 Ma (Jayananda et al., unpub. data).

Detrital zircon record from sedimentary sequences corresponding to different stratigraphic levels of Chitradurga basin yield wide range of U-Pb ages for their provenance. Krapez et al. (2020) documented older population of detrital zircon ages of 3386–2997 Ma (main population ca. 3161 ± 4 Ma) and younger detrital zircon population of 2610 ± 9 Ma from the basal unit of Talya conglomerate whilst detrital zircon ages of 3384–2990 Ma ages with main population 3276 ± 5 Ma and youngest age ca. 2597 ± 18 Ma from middle Talya conglomerate unit at the base of the Chitradurga basin along its western margin. On the other hand, detrital zircons from stromatolite stratigraphically overlying Talya conglomerate in the western part of the basin also yield wide range of U-Pb ages 3273–2960 Ma with a youngest grain of 2650 ± 38 Ma (Khelen et al., 2020) whilst detrital zircons from overlying Vanivilas Formation provide U-Pb ages 3376 Ma and 2768 Ma which coincides with ages of the surrounding TTG basement (present study; Ravindran et al., 2023) and Chitradurga volcanics (Hokada et al., 2013; Jayananda et al., 2013b) reflecting their origin from composite provenance. Further east detrital zircons from sandstones belong to higher stratigraphic levels provide diverse U-Pb ages of 2680 Ma to 2630 Ma with older ages of 3260 Ma and 2980 Ma (Hokada et al., 2013) wherein younger 2680–2630 Ma probably sourced from 2676 ± 20 Ma to 2677 ± 2 Ma pyroclastic felsic flows and potassic granitoids (Chadwick et al., 2007; Hokada et al., 2013; Jayananda et al., 2006, 2013b; Mohan et al., 2023) whilst older ages 3260 Ma and 2980 Ma sourced from TTG basement and Bukkapatna granite respectively (present study; Nasheeth et al., 2015; Mohan et al., 2023; Chardon et al., 2011; Ravindran et al., 2023). Further, detrital zircons from metagreywackes belong to overlying the Hiriyur Formation indicate bimodal U-Pb ages ranging 2697 Ma - 2677 Ma and 2635–2611 Ma with few grains yielding ages as old as 3400–3300 Ma (Nasheeth et al., 2016). The morphology of these detrital zircons including their well-preserved hexagonal crystal facies (Nasheeth et al., 2016) argue for short transport distance and probably sourced from adjacent volcanic arcs and basement granitoids (Nasheeth et al., 2015; Jayananda et al., 2013, 2020; Mohan et al., 2014, 2020).

East of the Chitradurga boundary shear zone, detrital zircon age record of Javagondanahalli sediments reveal younger U-Pb ages of 2587–2548 Ma (Nasheeth et al., 2016) which probably sourced from adjacent granitoids of Sira gneiss complex and sanukitoids (Chardon et al., 2011; Jayananda et al., 2020; present study). Detrital zircons from Aimangala conglomerate located at the eastern boundary of Chitradurga belt coinciding with boundary shear zone yield detrital zircon U-Pb ages of 2554 ± 7 Ma to 2549 ± 7 Ma (main population) whilst few zircons yield older bimodal ages of 2701–2635 Ma and 3405–3346 Ma (Krapez et al., 2020). These zircons probably sourced from surrounding Chitradurga felsic pyroclastic flows and diverse granitoids immediately to the east (Jayananda et al., 2006, Jayananda et al., 2013a, Jayananda et al., 2020, present study; Chardon et al., 2011; Nasheeth et al., 2016; Mohan et al., 2014, 2023). The youngest detrital ages of 2554–2549 Ma probably correspond to final assembly of central block with western block through westward convergence of intervening oceanic crust, lower crustal flow and formation of orogenic plateau (Chardon et al.,

2011; Jayananda et al., 2020).

In summary, published detrital zircon ages of the present studied corridor across the boundary shear zone from the Sargur Group, Bababudan Group, Chitradurga and Javagondanahalli belt including Aimangala conglomerate reveal their derivation from composite provenance involving Chitradurga volcanics and surrounding diverse granitoids which are in agreement with magmatic zircon record that show successive stage of crustal growth during ca. 3400–3300 Ma, 3200–3100 Ma, 3000–2950 Ma, 2700–2600 Ma, and 2560–2540 Ma. Detrital monazite ages reveal thermal events associated with reworking during 3100 Ma, 3000–2950 Ma, 2700 Ma, and 2500 Ma (Chardon et al., 2011; Hokada et al., 2013; Jayananda et al., 2013a; Jayananda et al., 2020; Kaempf et al., 2024, 2025).

12.3.3. Multistage crustal reworking, metamorphism and craton building

The Archean basement across the Chitradurga boundary shear zone involved in three major reworking events during ca. 3000–2950 Ma, ca. 2600 Ma and ca. 2500 Ma as revealed by U–Pb zircon ages and Nd isotopic data of potassic granites which spatially associated to metamorphic events (present study; Chadwick et al., 2007; Jayananda et al., 2006, 2019; Ao et al., 2021; Mohan et al., 2014). West of the boundary shear zone, Hosdurga, Bellur-Nagamangala and Bukkapatna potassic plutons emplaced during ca. 3000–2950 Ma (Jayananda et al., 2019; Chardon et al., 2011; Mohan et al., 2023; Ravindran et al., 2023). The emplacement of these plutons preceded by major crustal accretion through ca. 3200–3150 Ma komatiite-basalt volcanism and TTG-type granitoid formation (Ao et al., 2021; Chardon, 1997; Ravindran et al., 2023; Mohan et al., 2023; Maya et al., 2017; Tushipokla and Jayananda, 2013) in plume-arc settings. The emplacement of these plutons spatially and temporally associated with the ca. 2960 Ma high-pressure (11 kbar and 700 °C) metamorphism (Kaempf et al., 2025). Elemental and Nd isotopic data of these plutons reveal their origin through dehydration melting of lower crustal TTGs with minor juvenile input (present study; Jayananda et al., 2019). The associated high-pressure assemblages (garnet-kyanite, kyanite-muscovite) coupled with elemental characteristics and pressure sensitive elemental data reveal melt generation through dehydration melting of thickened lower crust. Seismic tomographic data also consistent with thickened Archean crust (42–51 km) in western Dharwar craton (Gupta et al., 2003). Such melting in the lower part of thickened crust implies collisional settings. Tectonic fabrics analysis of greenstones and adjacent granitoid basement to the west of Chitradurga belt show major crustal-scale shear zones along the Nagamangala, Banasandra, J.C. Pura and Ghattihosahalli belts. We speculate that these shear zones may represent reactivated suture zones where the discrete Paleoproterozoic terrains assembled through collision which caused thickening and lower crustal melting led to generation of ca. 3000–2960 Ma potassic granites and associated 2960 Ma high pressure metamorphism.

The ca. 2620 Ma potassic plutons intruded the Chitradurga-Javagondanahalli greenstone contact zone. The Chitradurga pluton located between two volcanic arcs (Chitradurga and Javagondanahalli) and bounded by boundary shear zone imply possible collision of two arcs through subduction and slab break off driven the melting of thickened crust. On the other hand, the Arsikere-Banavara plutons intrude the granitoid basement and associated Sargur Group Kalyadi-Banavara belt. Elemental data (high SiO₂, K₂O, Rb, Ba, REE (50–750 ppm) coupled with Nd isotope data [$\epsilon_{Nd(T)}$ -3.5 to -6.3] and pressure sensitive elemental ratios argue for melting of pre-existing thickened crust (7–15 kbar). The ca. 2600 Ma crustal reworking also spatially associated with ca. 2618 Ma metamorphic event (Kaempf et al., 2025). Consequently, west of the boundary shear zone two events of reworking events of thickened crust during ca. 3000–2950 Ma and ca. 2620 Ma generated high-potassic granites which spatially associated with ca. 2950 and ca. 2600 Ma metamorphic events associated with two stage cratonization prior to regional thermal event at 2500 Ma affected the whole Archean crust in the Dharwar craton.

East of boundary shear zone three reworking events documented ca. 3140 Ma, ca. 2630 Ma and ca. 2510 Ma (Jayananda et al., 2011, 2013a; Kaempf et al., 2024). These thermal events spatially temporally associated with juvenile magmatic accretion. The ca. 3140 Ma upper amphibolite thermal event ≥ 8 kbar and $T \leq 700^\circ\text{C}$ from metapelite from Pavagada (Jayananda et al., 2013b; Kaempf et al., 2024) and 20 km north of Challakere (see Fig. 1b) which immediately preceded by ca. 3230 Ma TTG-type granitoid accretion (Jayananda et al., 2013a; Chardon et al., 2011). In contrast, the ca. 2630 Ma event is a UHT metamorphic event (930 °C and 6.7 kbar, with limited decompression at 5.7 kbar at $\leq 810^\circ\text{C}$) spatially associated with major juvenile accretion of transitional TTGs during ca. 2650–2630 Ma (Jayananda et al., 2020; Kaempf et al., 2024; present study). Finally, the ca. 2510 Ma thermal event affected the entire Archean crust in the Dharwar craton which spatially associated with widespread sanukitoid magmatism, reworking and assembly of crustal blocks through westward convergence along the Chitradurga boundary shear zone (Jayananda et al., 2020).

12.3.4. Archean surface environments, oxygenation of ocean and emergence of biosphere

The interplay of geological, chemical and biological processes that drove the oxygenation of oceans-atmosphere are spatially linked to the emergence of the biosphere. Most studies of surface environments of the early Earth indicate highly reducing conditions prior to 2500 Ma (Planavsky and Mills, 2021 and references therein). The signatures of redox sensitive elements and isotope biomarkers (Fe, N and S) in the Archean sedimentary record provide invaluable insights into surface environments, redox recycling and metabolic pathways of microbial activity (Thomazo et al., 2009). Globally redox sensitive isotope tracers reveal a possibility of oxygen generation on ocean surface and minor free oxygen in atmosphere few hundred million years prior to the Great Oxidation Event (Planavsky and Mills, 2021 and references therein). The preserved Archean volcanic-sedimentary record in the studied corridor in the Dharwar craton across the Chitradurga boundary shear zone provide key archives for evaluating the redox conditions of the Archean surface environments, oxygenation of ocean-atmosphere, microbial activity and building of habitable continents of the early Earth.

The studied corridor preserves a large section of Archean continental crust forming a wide time window (ca. 3400–2500 Ma) corresponding to volcano-sedimentary record of Sargur Group and Dharwar Supergroup. These stratigraphic units west of Chitradurga boundary shear zone with well-preserved sedimentary assemblages record redox changes in ocean-atmosphere, sea floor hydrothermal plumbing systems, microbial activity and iron oxide precipitation in ocean through 3400–2600 Ma. The sedimentary record in the older Sargur Group include chert, carbonate, pelite and BIFs. The Bababudan Group contains conglomerate, quartzite, phyllite and BIFs whilst the Chitradurga Group preserve thick stromatolitic carbonates, greywacke-argillite-carbonaceous shales and BIFs. The Sargur Group chert and BIFs occasionally contain sulphide grains which points to anoxic environments during their deposition ca. 3300–3000 Ma wherein dissolved iron in ocean appears to be supplied through hydrothermal plumbing system (Mukherjee et al., 2012). The REE patterns of BIFs without any Ce anomaly coupled with positive Eu anomalies suggest anoxic conditions with dominant hydrothermal input which is agreement with other redox sensitive elements like Ce, Nd and Cr (Mukherjee et al., 2012). The sedimentary record of Bababudan Group (ca. 2900–2700 Ma Kibbanahalli belt) contains quartzite and thick BIFs. These BIFs show alternate bands of cherts and iron oxides with variations in their thickness. Such variations in the thickness of chert and iron oxides attributed to dynamics of depositional environments particularly fluctuating redox conditions (Hashizume et al., 2016). Their REE patterns with or without negative Ce anomaly coupled with other elemental ratios Ce, Nd, Cr suggest transient redox conditions with possible appearance of oxygen in the oceans (Mukherjee et al., 2012). This is in consistent with Fe and N isotope data of BIFs bands suggesting periodic fluctuations with minimum oxygen availability on ocean surface linking

production of organics microbial activity (Hashizume et al., 2016). The younger (2700–2600 Ma) Chitradurga Group preserve thick carbonate-greywacke-argillite-carbonaceous shales - BIFs sequences. The 2639±32 Ma carbonate sequence (Russell et al., 1996) forming spectacular stromatolite (see Fig. 3e) in the western part of the basin reflecting the microbial activity in the shelf environments with oxygen availability on the ocean surface. Furthermore, depleted oxygen isotope signatures (^{18}O -13.51 ‰ VPDB to -8.46 ‰ VPDB) attributed to higher sea water temperatures (Khelen et al., 2020) during their precipitation. The documented Ce anomalies either negative or absence of Ce anomalies in BIFs of the Chitradurga basin probably related to periodic fluctuations in redox conditions of ocean surface with minimum oxygen availability on the ocean surface (Mukherjee et al., 2025). The sulphur – Mass Independent Fractionation (S-MIF) study of sulphides from the Dharwar Supergroup Bababudan and Chitradurga basins reveal a progressive change in atmospheric chemistry from lower to upper stratigraphic levels through Archean (Mishima et al., 2017) with the first appearance of oxygen in the Neoproterozoic which increased significantly by the end of Archean.

12.4. Geodynamic evolution through time and formation of a habitable continent

12.4.1. Multistage accretional and collisional orogenesis

Published record combined with our new multi-disciplinary data on the crustal corridor across the Chitradurga boundary shear zone reveal two crustal blocks which preserve prolonged (ca. 3500–2500 Ma) crustal history of magmatism, reworking, surface environments, evolving tectonics of the Archean Earth. These crustal blocks formed independently through successive stages of crust formation, reworking in distinct geodynamic environments and finally assembled through collisional orogenesis (Jayananda et al., 2020).

Zircon U-Pb ages of TTG-type granitoids (Jayananda et al., 2015, 2023; Guitreau et al., 2017; Ranjan et al., 2020; Ao et al., 2021; Wang et al., 2023) and detrital zircon U-Pb ages reveal Archean basement in the western block formed during 3600–3200 Ma (Lancaster et al., 2015; Wang and Santosh, 2019; Mitra et al., 2023; Khelen et al., 2020; Mallens et al., 2023). The oldest (3600–3500 Ma) granitoids of a trondhjemite affinity formed by melting of the base of the oceanic plateau at shallow levels (Jayananda et al., 2015, 2023; Guitreau et al., 2017) followed by ca. 3300–3200 Ma TTG-type granitoids originated by melting of oceanic crust or oceanic arc crust at different crustal depths (Jayananda et al., 2015; Ranjan et al., 2020). The sub-contemporaneous adjacent ca. 3400–3200 Ma komatiite dominated greenstone volcanics formed in distinct tectonic settings including plume derived oceanic plateaus, oceanic arcs or accreted section of ocean ridge (Jayananda et al., 2008, 2016, 2023; Mukherjee et al., 2012; Tushipokla and Jayananda, 2013; Maya et al., 2017). Assembly of oceanic plateau, oceanic arc and oceanic ridge through convergence of intervening oceanic crust and eventual slab breakoff drive mantle upwelling which caused melting of juvenile crust/pre-existing TTGs at mantle depth resulting in generation of ca. 3200–3150 Ma hot diapiric trondhjemite magmas (Jayananda et al., 2023; Guitreau et al., 2017). Emplacement of hot trondhjemite magmas and associated mantle upwelling caused partial convective overturn of middle to lower crust driving gravitational collapse of high-density greenstones instabilities leading to development of dome and keel structures and stabilization of Archean crust close to 3100 Ma (Jayananda et al., 2023).

On the eroded stabilized Archean basement, the ca. 3000 Ma intra-cratonic volcano-sedimentary basins including Bababudan Group Kibbanahalli arm, Yediyur, western margin of Chitradurga belt were deposited. These basins contain voluminous amygdaloidal basalts to felsic pyroclastic flows on the higher stratigraphic levels revealing their origin possibly related hotspot arrival beneath the stabilized granitoid basement causing opening of basins. Strain fabrics mapping and kinematic analysis are compatible with a sagducting trough (Chardon et al.,

1996, 1998) possibly linked to hotspot, whilst other workers attributed development of these basins to failed rift settings (e.g., Sreehari et al., 2021, 2024).

The Chitradurga basin (Seshadri et al., 1981) surrounding basement form dome-basin architecture (Jayananda et al., 2006) whilst other workers attributed to fold interference (Naqvi, 1973; Mukhopadhyay et al., 1981; Mukhopadhyay and Baral, 1985; Naha et al., 1995). The tectonic context of basin development was attributed to a back arc setting (Chadwick et al., 1981) whilst more recent combined field based stratigraphic, fabric data and U-Pb zircon data argue for two stage back arc basins behind an east facing continental arc (Krapez et al., 2020). The volcano-sedimentary assemblages of the Chitradurga belt bounded by older continental crust on either side compatible with a back-arc basin. The matrix supported Talya conglomerate in the western part of the Chitradurga basin interpreted to be glaciomarine origin (Ojakangas et al., 2014) wherein the clasts were sourced from pre-existing granite-greenstone terrain as well as raising volcanic arcs of the Chitradurga basin as revealed by detrital zircon ages of 3386–2597 Ma (Krapez et al., 2020). The absence of clast supported framework in Kurumardikere (K. M. Kere) conglomerate in the central part (see Fig. 3c) with dominant chert fragments embedded in clay/silt infer fragmentation within oceanic environment which compatible with rifting. The Talya and Kurumardikere conglomerates separating the different stratigraphic units (Seshadri et al., 1981; Chadwick et al., 1981; Krapez et al., 2020; present study) containing volcanic-sedimentary assemblages such as sandstones, stromatolitic carbonates, basalts, greywacke-argillite-BIFs together with major high-strain zones containing high pressure assemblages infer a composite back arc scenario involving initial rifting, drifting and convergence. The distinct volcano-sedimentary units on either side of the rift related conglomerate (Kurumardikere) suggest at least two distinct volcanic arcs with intervening basins in the back arc setting. The volcanic arcs in the west containing voluminous mafic rocks associated with shallow water shelf sedimentary assemblages like conglomerate, arenite and stromatolite-carbonate indicate continental arc signatures whilst in the central part oceanic conglomerates depict initiation of rift in oceanic setting. Further towards east volcanic arcs associated with deep water turbidite sequences like greywacke-argillite-carbonaceous shales- BIFs suggests deepest conditions. The distribution of shallow water sedimentary assemblages in the west can be attributed to detritus sourced from older continental crust in shelf conditions whilst deep water turbidite sequences (greywacke-argillite-carbonaceous shale-BIFs) to the east sourced from mixed provenance comprising juvenile volcanic arcs (Javagondanahalli belt; present study; Krapez et al., 2020) and pre-existing crust (Chardon et al., 2011; Nasheeth et al., 2015, 2016). The development of Chitradurga back-arc architecture attributed to westward convergence of oceanic plate and eventual slab breakoff beneath matured continental crust (Jayananda et al., 2020) leading to mantle upwelling forms driving force for rifting, subsidence and drifting of oceanic crust.

East of boundary shear zone (see Fig. 1b) the ca. 2740 Ma Javagondanahalli greenstone belt in the north and Kunigal belt in the south is dominated by voluminous mafic volcanics, minor tuffs with thin carbonate or chert layers and BIFs. Lithological sequence, metamorphic assemblages (low to high pressure) with clockwise P-T paths coupled with mix of voluminous ca. 2690–2600 Ma juvenile transitional TTGs and large disrupted boudins of ca. 3300–3170 Ma migmatitic TTGs with quartzite-pelite which involved in low pressure metamorphism. These lithological characteristics together with shallow fabrics transforming into steep fabrics showing SW verging lineation from basement to the greenstone belt contacts suggest westward convergence of oceanic lithosphere beneath the pre-existing continental crust (Chadwick et al., 2000; Jayananda et al., 2020).

East of Javagondanahalli greenstone belt the magmatic protoliths of transitional TTGs were emplaced ca. 2670–2600 Ma (present study; Jayananda et al., 2020). Petrologic data including clockwise metamorphic P-T-t paths combined with elemental (high Mg, Ca, K, high REE

and isotope data [$\epsilon_{\text{Hf}(T)} = +0.67$ to $+5.61$; $\epsilon_{\text{Nd}(T)} = +0.52$ to $+4.23$] consistent with a composite source involving major mantle input and newly formed arc crust in their genesis in subduction zone settings (Jayananda et al., 2020; Kaempfer et al., 2024; present study). Collision of Javagondanahalli and Chitradurga volcanic arcs through westward convergence of intervening oceanic crust caused crustal thickening and lower crustal melting generated potassic Chitradurga and Jampalnai-kankote granites. Continued subduction and eventual slab breakoff beneath the western block led to mantle upwelling that drives lower crustal melting producing potassic granites (ca. 2620 Ma Arsikere-Banavara granites) and also opening of rift system further west along the eastern margin of Shimoga basin.

Further east of Javagondanahalli greenstone belt, high-pressure (11–14 kbar) garnet bearing ultramafic rock from NW trending shear zone indicate in-situ Lu-Hf garnet age of 2617 Ma (Kaempfer et al., 2024) which coincides with the emplacement age of ca. 2615 Ma Chitradurga granite, ca. 2600 Ma Jampalnai-kankote granite. The high-pressure garnet-bearing ultramafic rocks close to eclogite facies could represent slices of abducted oceanic crust from mantle depth during the collision of Chitradurga and Javagondanahalli arcs through westward convergence. The youngest detrital zircon ages of ca. 2540 Ma documented in the eastern margin of Chitradurga basin (Krapez et al., 2020) reveal closure of basin through collision of tectonic units in the context of westward convergence of intervening oceanic crust. This is consistent with the seismic reflection profiles suggesting westward convergence of the oceanic lithosphere along the Chitradurga boundary shear zone (Vijaya Rao et al., 2015; Mandal et al., 2018). The final stage of collision of two crustal blocks along the Chitradurga boundary shear zone is marked by the intrusion of aplitic Maddakkanahalli granite at 2507 ± 5 Ma (Chardon et al., 2011) which led to final stabilization of the Archean crust into habitable continent marking the end of the Archean.

13. Conclusions

The salient conclusions of this study are summarized as follows:

1. The two crustal domains across the Chitradurga boundary shear zone show distinct lithologic assemblages, crustal architecture, magmatic and metamorphic records and crust accretion histories. The western domain preserves older basement with ca. 3400–3200 Ma TTGs-komatiite assemblages, ca. 3000 Ma and ca. 2600 Ma potassic granites and sediment-rich greenstone sequences whilst the eastern domain is dominated by voluminous 2676–2600 Ma transitional TTGs with remnants of ca. 3300–3170 Ma TTGs, and voluminous juvenile intrusions of sanukitoid affinity. Furthermore, thick crust (43–45 km) and world class dome-keel structures are bounded by widely spaced shear zones to the west whilst thin crust (30–33 km) with flat fabrics linked to lower crustal flow and closely spaced regional interfering shear zone pattern to the east of the boundary shear zone.
2. The western crustal domain preserves two moderate to high pressure metamorphic events ca. 3100–3000 Ma and ca. 2500 Ma whilst the crustal domain to the east of the boundary shear zone shows moderate to low pressure and moderate to ultra-high temperature events at ca. 3140 Ma, 2630 Ma and 2500 Ma. The Chitradurga boundary shear zone itself contains ca. 2630 Ma high pressure (12–14 kbar, 400–700°C) and ca. 2500 Ma high temperature (ca. 800°C) assemblages with pressure (7–11 kbar) linked to two distinct tectonic events. Zircon U-Pb ages of magmatic record and detrital zircon coupled with whole-rock Nd isotopes and in-situ Hf isotope data of zircon show successive stages of crust accretion and continental growth in the western domain at ca. 3600–3500 Ma, 3400–3300 Ma, 3200–3100 Ma whilst the eastern domain shows widespread 2700–2500 Ma accretion with high rates of crustal growth.
3. The incompatible element data combined with whole-rock Nd and zircon in-situ Hf isotope data reveal high rates of crustal growth and

evolution of complimentary shallow to deep mantle reservoirs during ca. 3400–3300 Ma, 3200–3100 Ma, 3000–2850 Ma, and 2700–2500 Ma plume-arc settings. The existence of deep to shallow depleted mantle at 3400 Ma imply an earlier event of mantle differentiation during 3600–3700 Ma perhaps even Hadean as revealed by ^{142}Nd anomalies in Sargur Group volcanics. The enrichment of shallow mantle reservoirs during 2700–2560 Ma can be attributed to sediment input into subduction zone by weathering of rising magmatic arcs. Two major episodes of reworking of thickened lower crust in the western domain in the context of collision lead to emplacement of potassic plutons during ca. 3000–2950 Ma and 2620–2600 Ma and stabilization of crust. In contrast, one major reworking in the eastern domain close to 2520–2500 Ma resulted in anatectic granites and 2507 Ma old Maddakkanahalli granite spatially linked to the final collision of two crustal blocks along the Chitradurga boundary shear zone.

4. Tectonic fabrics data coupled with timing as well as metamorphic P-T conditions, incompatible elemental data show that the preserved old (>3300 Ma) crust in the western domain originated in plume-arc setting. Collision of volcanic arcs and/or micro-blocks with eventual slab break off caused melting of thickened crust resulted in the emplacement of potassic plutons ca. 3000–2950 Ma 2620 Ma. The ca. 3000–2850 Ma intracratonic greenstone basins developed through sagduction driven slab break off generated hotspots whilst 2700–2600 Ma greenstone basins developed in arc settings (back-arc to fore arc). Final collision of two crustal blocks along the Chitradurga boundary shear zone with eventual slab break off led to mantle upwelling causing widespread mantle melting, widespread juvenile magma emplacement, reworking, lower crustal flow, hot metamorphism and final cratonization leading to establishment of large stable continental land mass.
5. Field evidence for microbial activity together with published redox sensitive elements and isotope biomarkers (Fe, N, S-MIF, C, O) show a fluctuating redox conditions in ocean - atmosphere and transition from anoxic to oxygenated environments and emergence of biosphere culminating with the formation of habitable continents at the dawn of Great Oxidation Event.

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CRedit authorship contribution

M. Jayananda – Designing project, funding acquisition, conceptualization, investigation and writing of original manuscript.

Aadhisheshan –Field work, data compilation, curation and processing, formal data analysis.

Martin Guitreau – LA-ICP-MS U-Pb zircon and LA-MC-ICP-MS Lu-Hf isotope analysis, validation, intellectual inputs.

S.V. Balaji Manasa Rao- Field work, zircon separation, formal data analysis.

Pritam Nasipuri – Mineral chemistry, validation, formal data analysis.

M. Satyanarayanan - Trace element analysis, data curation and validation.

M. Owada – Major element analysis, data validation and intellectual inputs.

Declaration of competing interest

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

M. Jayananda reports was provided by University of Hyderabad. M. Jayananda reports a relationship with University of Hyderabad that includes: employment and non-financial support. If there are other authors, they declare that they have no known competing financial

interests or personal relationships that could have appeared to influence the work reported in this paper.

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Data availability

Data will be made available on request.

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