Paleoproterozoic (~ 1.88 – 1.89 Ga) ultramafic–mafic sills, Cuddapah basin, India—revisited: Implications for interaction between mantle plume and metasomatized subcontinental lithospheric mantle

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A number of mafic–ultramafic sills, which are supposed to be part of the widespread ~ 1.88 – 1.89 Ga Hampi–Bastanar Large Igneous Province of the Indian shield, are reported to intrude into sedimentary/meta-sedimentary rocks of the Proterozoic Cuddapah basin. Mineral chemistry of chrome-spinel, olivine and pyroxene from the cumulate ultramafic rocks of the Pulivendla sills, emplaced at the base of Tadpatri Formation of the Cuddapah basin, is presented for better understanding on their nature and genesis. The intermediate Cr#, low Mg#, high Fe^{3+} and Ti (apfu) coupled with relatively low Al (apfu) content of studied spinel indicate their Alaskan-type nature. Moreover, the Ca-rich clinopyroxene together with classic variations of Al_2O_3 , TiO_2 and SiO_2 follow the trend of arc type of intrusions. The calculated Al_2O_3 content of parental melt of the studied rocks also suggests arc-type characteristics. Whole-rock geochemistry show fractionated chondrite-normalized REE patterns, higher Ba/Nb and low Nb/La that suggest contribution from an enriched mantle source, whereas higher Th/Yb and negative Ta–Nb–Ti anomaly on the primitive mantle-normalized multi-element spidergrams emphasize the involvement of a subducted component in the lithospheric mantle source. Although, the mineral chemistry and geochemistry is akin to Alaskan-type intrusion, lack of concentric zoning of lithologies (olivine-rich ultramafic rocks in centre surrounded by mafic rocks), rarity of primary hornblende and abundance of orthopyroxene contradict it to be an Alaskan-type intrusion. It is suggested that the mantle source region of the Pulivendla ultramafic–mafic rocks was modified by a fluid-induced metasomatism during an ancient subduction event. Further, Al-in-olivine thermometry suggests crystallization temperature between 1410° and 1484°C , which is 260°C higher than the average temperature of MORB. Similarly, the estimated mantle potential temperature is also significantly higher ($>1600^\circ\text{C}$) than the different secular cooling models of the earth at ~ 1.88 – 1.89 Ga and consistent with the thermal regime of a mantle plume. The existence of mantle plume during 1.88 – 1.90 Ga, which has played an important role in the genesis of mafic–ultramafic sills of the Cuddapah basin, is also well supported by a radiating mafic dyke swarm, domal uplift and magmatic underplating.

Keywords. Alaskan-type rock; Al-in-olivine thermometry; subduction; SCLM; mantle plume.

1. Introduction

The advent of a mantle plume head is perceived by thermal anomalies in the upper mantle, which may be hundreds of degrees hotter than the ambient mantle (e.g., Wolfe *et al.* 1997; Bijwaard and Spakman 1999; Cao *et al.* 2011; Cibrowski *et al.* 2017). A number of geological, geochemical, and geophysical criteria can be exercised to identify mantle plume and associated magmatic provinces in the geological record (e.g., Ernst and Buchan 2003; Campbell 2005; Ernst 2014); however, integration of crystallization and mantle potential temperature would provide compelling evidence to decipher the role of a mantle plume. The study of temperature dependent behaviour of Al in some key minerals, such as chrome-spinel and olivine (Al-in-olivine thermometer) in mafic and ultramafic rocks, provides an opportunity to assess the maximum crystallization temperature (Coogan *et al.* 2014), which can also be approximated towards the corresponding mantle potential temperature (T_p). This thermometer also successfully explains the standard plume model for several Large Igneous Provinces around the globe (Coogan *et al.* 2014; Heinonen *et al.* 2015; Xu and Liu 2016).

It is known that spinel and olivine are formed at the earliest stage of crystallization of a primitive mantle-derived magma; therefore provide information on initial crystallization conditions (Irvine 1965, 1967; Pichavant and Macdonald 2007). Spinel, particularly chrome-spinel, is also sensitive to chemical and thermal conditions of a crystallizing rock and its composition is controlled by multiple factors such as source compositions, oxygen fugacity, crystallization temperature, cooling rate and pressure (Kamenetsky *et al.* 2001); therefore, it is widely used to explicate petrogenetic aspects (Irvine 1965, 1967; Dick and Bullen 1984).

A number of 1.88–1.89 Ga mafic/ultramafic sills are widely intruded and well-exposed in the lower stratigraphic horizons of the Proterozoic Cuddapah intracratonic basin (Anand *et al.* 2003; French *et al.* 2008) (see figure 1). It is suggested that these mafic/ultramafic sills are integral part of a widespread ~1.88–1.89 Ga Large Igneous Province (LIP) (Ernst and Srivastava 2008; French *et al.* 2008; Belica *et al.* 2014; Srivastava *et al.* 2014a; Srivastava and Samal 2019), identified as a plume derived Hampi–Bastanar LIP (Samal *et al.* 2019). However, Sheppard *et al.* (2017) have provided a range of emplacement ages (1864 ± 13 and 1858 ± 16 Ma) for the felsic tuffs from the upper part of the

Tadpatri Formation suggesting a series of magmatic events for ~30 Ma and argued for a non-plume origin. This is also substantiated by Anand *et al.* (2003) who have suggested that the melts of Pulivendla sills was generated in a mantle potential temperature of ~1500°C below the Cuddapah Basin which is consistent with the secular cooling model of the mantle.

This work is aimed to provide testimony to the genesis of Pulivendla mafic–ultramafic sills; whether it is derived from plume induced mantle melts or some other processes involved. A detailed mineralogical study has been carried out on chrome-spinel, olivine and pyroxene to assess the nature of parental melt, mantle source, melting conditions and geodynamic setting. The new mineralogical data together with available geochemical data (Anand *et al.* 2003) are examined to provide a broad picture on nature of subcontinental lithospheric mantle (SCLM) below the eastern Dharwar craton and Cuddapah basin at ~1.88–1.89 Ga.

2. Geological setting

The Indian peninsula, comprising a number of Archean cratonic blocks, mainly consists of greenstone and high-grade metamorphic belts embedded in tonalite–trondhjemite gneissic complexes with Archean protoliths and ~2.5 Ga granitic activities (Naqvi and Rogers 1987; Mahadevan 2008; Ramakrishnan and Vaidyanadhan 2010) and later, during Paleoproterozoic, transected by several mafic dyke swarms (Murthy 1987; Srivastava *et al.* 2008), which are thought to be part of different LIPs (Söderlund *et al.* 2019; Samal *et al.* 2019). Most of these cratonic blocks also witnessed development of several intracratonic basins, designated as Purana basins, in the late Palaeoproterozoic to Neoproterozoic (Radhakrishna 1987; Chakraborti *et al.* 2010; Kale 2016). Among these Purana basins of the Indian peninsula, Cuddapah basin, occurring along the eastern margin of the Dharwar craton, is the second largest and the most studied Proterozoic basin (cf. Nagaraja Rao *et al.* 1987; Patranabis-Deb *et al.* 2012; Saha and Tripathy 2012; Matin 2015). The crescent shaped, easterly concave, N–S trending Cuddapah basin (extending over a length of about 450 km and a width of 200 km and covering an area of 44,500 km²) comprises 6–12 km thick successions of sedimentary rocks, intercalated with igneous rocks (Kale 1991; Nagaraja Rao *et al.* 1987; Ramakrishnan and Vaidyanadhan 2010). The

Figure 1. Geological map of the Cuddapah Basin (modified after Chakraborty *et al.* 2016) and surroundings (modified after Samal *et al.* 2019). (*) locations of dated samples; ^{40}Ar – ^{39}Ar phlogopite age (Anand *et al.* 2003) and U–Pb IDTIMS baddeleyite age (French *et al.* 2008).

Cuddapah Supergroup consists of unconformity-bound successions of Papaghni, Chitravati and Nallamalai Groups and Srisailem Quartzite Formation (Nagaraja Rao *et al.* 1987; Matin 2015) (see figure 1). Additionally, on the basis of sediment characteristics, spatial distribution and evolution of the basin through time, four sub-basins within the Cuddapah Basin, viz., Papaghni, Kurnool, Srisailem and Palnad sub-basins, have been identified (cf. Nagaraja Rao *et al.* 1987; Patranabis-Deb *et al.* 2012; Saha and Tripathy 2012; Matin 2015).

Both extrusive and intrusive magmatic rocks are well-exposed in the older formations of the Cuddapah Supergroup (Nagaraja Rao *et al.* 1987). Recently, based on geographical locations and nature of occurrence in various stratigraphic horizons, Sesha Sai *et al.* (2017) have reviewed the Paleoproterozoic magmatic events of the Cuddapah basin. The Vempalle Formation of Papaghni Group as well as younger Tadpatri Formation of Chitravati Group is intruded by several mafic–ultramafic sills (cf., Patranabis-Deb *et al.* 2012; Saha

and Tripathy 2012; Chakraborty *et al.* 2016; Sesha Sai *et al.* 2017). The thickness of these sills is variable in different stratigraphic successions; ~200 m in the lower Tadpatri Formation whereas ~20 m in the other younger sequences (Anand *et al.* 2003). Mafic-ultramafic sills of the present study are part of the Tadpatri Formation, distributed over the southwestern margin of the basin, named as Pulivendla sills (see figure 1); it is a differentiated body represented by a coarse-grained ultramafic rock at the base and gabbro in the upper sequence (figure 2). This sill is dated by $^{40}\text{Ar}/^{39}\text{Ar}$ (phlogopite age; Anand *et al.* 2003) as well as U–Pb IDTIMS (baddeleyite age; French *et al.* 2008) methods, which suggest their emplacement at 1899 ± 20 and 1885.4 ± 3.1 Ma, respectively. Besides this differentiated sill, some dolerite sills (called as minor sills by Anand *et al.* 2003) are also reported in the Tadpatri Formation.

A number of Paleoproterozoic mafic magmatic events, mostly in form of dyke swarms, are reported all around the Cuddapah basin (Ernst and Srivastava 2008; French and Heaman 2010; Söderlund *et al.* 2019; Samal *et al.* 2019). Amongst these, ~1.88–1.89 Ga Hampi–Bastanar LIP (Samal *et al.* 2019), which spread almost in entire Indian shield covering the Dharwar craton (Hampi swarm; Halls *et al.* 2007), Cuddapah basin (as Pulivendla sills; Anand *et al.* 2003; French *et al.* 2008), Bastar craton (Bastanar swarm; French *et al.* 2008; Shellnut *et al.* 2018; Samal *et al.* 2019), and Himalayan Mountain Range (Srivastava and Samal 2019), is particular interest of the present study.

3. Instrumental technique

All mineral analyses are carried out in the Electron Probe Micro Analyzer (EPMA) CAMECA SX-Five instrument at the DST-SERB National Facility, Department of Geology, Banaras Hindu

University, India. The instrument with SXFive Software at a voltage of 15 kV and current 40 nA with a LaB6 source in the electron gun for generation of electron beam is used during this work. Natural silicate mineral, i.e., andardite is used as internal standard to verify positions of crystals (SP1-TAP, SP2-LiF, SP3-LPET, SP4-LTAP and SP5-PET) with respect to corresponding wavelength dispersive (WD) spectrometers (SP#) in CAMECA SX-Five instrument. Routine calibration, acquisition, quantification and data processing were carried out using SxSAB version 6.1 and SX-Results softwares of CAMECA. The precision of the analysis is better than 1% for major element oxides and 5% for trace elements from the repeated analysis of standards. Major and trace elements in olivine were analysed at an acceleration voltage of 15 kV, beam current of 40 nA and the calibration setting is given in table 1. All the generated data are presented in tables 2, 3 and 4. All Fe are considered as Fe^{2+} in silicates, and Fe^{2+} and Fe^{3+} are calculated for spinel assuming stoichiometry.

4. Results

4.1 Petrography and mineral chemistry

The samples of the present study are collected from the ultramafic–mafic sills, exposed in and around Pulivendla area and are marked in figure 1. The samples are fresh, coarse grained and chiefly consists of cumulus olivine and clinopyroxene. Orthopyroxene, phlogopite and spinel are common accessory phases. Few fine-grained plagioclase feldspars (<5 mode %) are also noticed; which are calcic in nature crystallizing from inter-cumulus liquid. Similarly, the phlogopite crystals are also crystallized from the inter-cumulus liquid. Subhedral to euhedral spinel crystals are observed as inclusions within and interstitial areas of subhedral

Figure 2. Field photograph of Pulivendla sill; the inside photograph is larger version showing ultramafic part of the sill.

Table 1. Calibration settings for major and trace elements of olivine.

X-ray line	Crystal	Peak time	Background (–ve)	Background (+ve)	Standard	Formula	Detection limit (in ppm)	Std. dev. (in %)
Mg K α	TAP	30	500	500	Periclase	MgO	173	0.1102
Al K α	LTAP	240	500	500	Corundum	Al ₂ O ₃	22	0.0022
Si K α	TAP	30	500	500	Wollastonite	CaSiO ₃	102	0.0678
Ca K α	LPET	240	500	500	Wollastonite	CaSiO ₃	55	0.0055
Ti K α	LPET	120	500	500	Rutile	TiO ₂	61	0.0052
Mn K α	LIF	120	500	500	Rhodonite	(Mn, Fe, Mg, Ca)SiO ₃	22	0.0287
Fe K α	LIF	120	500	500	Hematite	Fe ₂ O ₃	116	0.1684
Na K α	LTAP	120	500	500	Halite	NaCl	38	0.0038
Cr K α	LPET	240	500	500	Chromite	Cr ₂ O ₃	69	0.0082
Ni K α	LIF	240	500	500	Nickel	Ni	438	0.0420

Table 2. Representative EPMA data (wt%) of olivine from Pulivendla ultramafic rocks.

Oxide wt%	1	2	3	4	5	6	7	8	9	10
SiO ₂	39.38	39.48	39.66	39.73	39.43	39.30	39.14	39.30	39.04	39.31
TiO ₂	0.02	0.02	0.02	0.03	0.04	0.03	0.03	0.02	0.03	0.03
Al ₂ O ₃	0.04	0.04	0.03	0.02	0.02	0.02	0.02	0.04	0.02	0.02
Cr ₂ O ₃	0.03	0.03	0.03	0.03	0.03	0.02	0.03	0.03	0.03	0.02
FeO	18.26	18.27	18.27	18.57	18.72	18.63	19.01	18.68	18.96	18.70
MnO	0.24	0.25	0.25	0.25	0.25	0.27	0.24	0.26	0.27	0.26
MgO	41.69	41.59	41.63	41.42	41.40	41.61	41.23	41.31	41.43	41.43
NiO	0.20	0.21	0.20	0.24	0.19	0.22	0.21	0.22	0.20	0.21
CaO	0.16	0.16	0.16	0.13	0.09	0.10	0.12	0.14	0.10	0.10
Total	100.02	100.03	100.25	100.42	100.18	100.18	100.03	100.00	100.08	100.09
Cations based on 4 oxygen										
Si	1.005	1.008	1.009	1.010	1.006	1.003	1.003	1.006	1.000	1.005
Ti	0.000	0.000	0.000	0.001	0.001	0.001	0.001	0.000	0.001	0.001
Al	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001
Cr	0.001	0.001	0.001	0.001	0.001	0.000	0.001	0.001	0.001	0.000
Fe(ii)	0.390	0.390	0.389	0.395	0.399	0.398	0.407	0.400	0.406	0.400
Mn	0.005	0.005	0.005	0.005	0.005	0.006	0.005	0.006	0.006	0.006
Mg	1.587	1.582	1.580	1.571	1.575	1.583	1.575	1.576	1.581	1.578
Ni	0.004	0.004	0.004	0.005	0.004	0.004	0.004	0.004	0.004	0.004
Ca	0.004	0.004	0.004	0.004	0.003	0.003	0.003	0.004	0.003	0.003
Fo	80.06	80.02	80.02	79.69	79.55	79.70	79.24	79.54	79.35	79.57
Fa	19.67	19.71	19.70	20.04	20.17	20.01	20.50	20.17	20.36	20.14
Tp	0.26	0.27	0.28	0.28	0.27	0.29	0.27	0.28	0.29	0.29
Mg#	80.27	80.23	80.25	79.91	79.77	79.93	79.45	79.77	79.58	79.80
Trace elements (in ppm)										
Ca	1164	1173	1146	921	671	728	840	1011	742	744
Al	186	186	174	130	123	98	122	200	103	112
Ni	1594	1637	1594	1895	1525	1710	1663	1694	1587	1666
Ti	127	113	125	168	210	165	167	140	189	179
Cr	221	198	200	199	188	124	178	174	237	163
Mn	1860	1908	1955	1955	1950	2062	1886	1993	2054	2030
(Feo/ MgO) _{melt}	1.46	1.46	1.46	1.49	1.51	1.49	1.54	1.51	1.53	1.50

Table 3. Representative EPMA data (wt%) of chrome spinels from Pulivendla ultramafic rocks.

Oxide wt%	1	2	3	4	5	6	7	8	9	10
SiO ₂	0.00	0.00	0.28	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TiO ₂	2.73	2.82	7.64	4.45	1.79	1.60	1.79	6.17	6.26	4.32
Al ₂ O ₃	11.84	8.99	6.10	9.12	13.39	13.63	13.24	7.67	7.98	9.25
Cr ₂ O ₃	38.80	38.07	31.11	36.61	41.07	41.41	41.11	32.22	32.06	36.21
Fe ₂ O ₃	14.12	16.65	16.94	15.26	12.22	11.94	13.00	17.36	17.16	15.34
FeO	24.21	27.25	31.34	26.89	22.96	22.79	22.81	30.01	29.82	26.58
MnO	0.46	0.33	0.52	0.52	0.44	0.27	0.42	0.48	0.42	0.45
MgO	7.36	5.18	5.09	6.20	7.85	7.97	8.12	5.03	5.28	6.22
CaO	0.13	0.06	0.09	0.10	0.12	0.13	0.09	0.07	0.07	0.14
TOTAL	99.64	99.34	99.12	99.15	99.83	99.74	100.57	99.01	99.06	98.51
Cations based on 32 oxygens										
Si	0.000	0.000	0.080	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ti	0.553	0.591	1.625	0.926	0.357	0.320	0.355	1.306	1.320	1.550
Al	3.760	2.950	2.033	2.971	4.199	4.270	4.120	2.544	2.635	1.856
Cr	8.268	8.380	6.954	8.003	8.640	8.702	8.586	7.168	7.106	6.614
Fe(iii)	2.865	3.489	3.603	3.175	2.446	2.388	2.583	3.676	3.620	4.429
Fe(ii)	5.458	6.344	7.408	6.217	5.108	5.065	5.038	7.061	6.990	8.233
Mn	0.104	0.077	0.125	0.123	0.099	0.061	0.093	0.116	0.100	0.085
Mg	2.956	2.152	2.146	2.556	3.115	3.159	3.198	2.110	2.208	1.206
Ca	0.036	0.017	0.026	0.029	0.035	0.036	0.026	0.020	0.022	0.027
Total	24.000	24.000	24.000	24.000	24.000	24.000	24.000	24.000	24.000	24.000
Cr#	68.74	73.97	77.38	72.92	67.29	67.08	67.57	73.80	72.95	78.09
Fe ³⁺ #	19.23	23.54	28.62	22.44	16.00	15.55	16.90	27.46	27.09	34.33
Mg#	35.13	25.33	22.46	29.14	37.88	38.41	38.83	23.00	24.00	12.77
(Al ₂ O ₃) _{melt}	11.09	9.90	8.44	9.96	11.67	11.76	11.62	9.27	9.42	10.02
(TiO ₂) _{melt}	3.06	3.16	8.42	4.94	2.03	1.83	2.04	6.81	6.91	4.80
(FeO/MgO) _{melt}	2.38	3.50	3.81	2.94	2.21	2.18	2.10	3.83	3.66	2.50

olivine grains (figure 3). Poikilitic texture, olivine grains embedded within pyroxene grains (figure 3), is common in these studied samples. Based on petrography, mineral chemistry and geochemistry, Anand *et al.* (2003) have classified these ultramafic rocks as picrite, whereas, based on modal mineralogy, Sesha Sai *et al.* (2017) have identified them as plagioclase bearing ultramafic rock. Based on observed mineralogy and texture, the studied rock type is classified as plagioclase-bearing ultramafic rock (Le Maitre 2002).

Olivine compositions show a very limited variation in forsterite (Fo) contents; vary from Fo₇₇ to Fo₈₀ (table 2). NiO contents vary from 0.19 to 0.25 and show positive correlation with Fo values. Trace element composition of olivine (being the first silicate mineral to crystallize from any primary melts) may act as a tracer of melting and crystallization history of magma. Therefore, some key trace elements (Ca, Cr, Al, and Ti) are also analyzed during this work. Ti, Al and Ca contents vary between 83 and 210, 98 and 204, and 671 and 1173 ppm, respectively. Spinel compositions

[Cr₂O₃ (~31–41 wt%), Al₂O₃ (~6–14 wt%), Fe₂O₃ (~12–17 wt%), and MgO (5.02–8.12 wt%)] suggest their chrome-spinel nature (see table 3). They have moderate Cr# (=Cr/[Cr + Al]) (0.67–0.78), low Mg# (=Mg/[Mg + Fe²⁺]) (0.12–0.38), and high TiO₂ (1.60–7.64 wt%) suggesting their derivation from a Ti-rich parental melt. Fe³⁺# (Fe³⁺/Fe³⁺+Al+Cr) of analyzed spinels is calculated stoichiometrically that ranges from 15 to 34. Most of the analyzed clinopyroxenes show augitic characteristics with few grains having diopsidic nature (Morimoto 1989); suggesting their Ca-rich nature and show limited variation in their composition (En₄₄₋₅₅ Fs₈₋₁₂ Wo₃₄₋₄₄) with Mg# (78–87), TiO₂ (0.19–1.42 wt%), Al₂O₃ (1.64–4.15 wt%), and Cr₂O₃ (0.58–1.24 wt%) (see table 4).

4.2 Estimation of crystallization temperature

The partitioning of ‘Al’ in forsterite-rich olivine and chrome-spinel is temperature dependent, therefore, Al-in-olivine thermometry, which is independent on parental melt composition or mineral-melt

Table 4. Representative EPMA data (wt%) of clinopyroxene from Pulivendla ultramafic rocks.

Oxide wt%	1	2	3	4	5	6	7	8	9	10
SiO ₂	51.87	49.63	52.27	50.52	51.38	52.16	50.53	51.97	52.16	51.86
TiO ₂	0.75	1.42	0.28	0.19	0.26	0.34	0.26	0.56	0.80	0.81
Al ₂ O ₃	2.66	4.15	1.82	1.88	1.97	1.64	1.85	2.38	2.47	2.27
Cr ₂ O ₃	0.84	0.73	0.58	0.76	0.87	0.77	0.74	1.24	1.15	0.82
FeO	6.17	5.49	5.52	5.59	5.28	5.36	8.12	5.21	5.95	5.89
MnO	0.06	0.19	0.04	0.12	0.22	0.17	0.25	0.19	0.23	0.19
MgO	16.84	15.61	19.39	19.07	18.53	20.97	16.78	16.51	16.13	16.38
CaO	20.89	21.44	19.48	20.56	20.12	18.27	20.70	20.22	20.85	20.11
Na ₂ O	0.27	0.40	0.42	0.40	0.26	0.30	0.19	0.47	0.46	0.47
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.01	0.02	0.01
Total	100.36	99.06	99.80	99.09	98.90	99.99	99.44	98.77	100.22	98.82
Cations based on 6 oxygen atoms										
Si	1.903	1.850	1.916	1.881	1.906	1.904	1.895	1.928	1.917	1.927
Ti	0.021	0.040	0.008	0.005	0.007	0.009	0.007	0.016	0.022	0.023
Al	0.115	0.182	0.079	0.083	0.086	0.070	0.082	0.104	0.107	0.099
Cr	0.024	0.021	0.017	0.022	0.026	0.022	0.022	0.036	0.033	0.024
Fe(ii)	0.189	0.171	0.169	0.174	0.164	0.163	0.255	0.162	0.183	0.183
Mn	0.002	0.006	0.001	0.004	0.007	0.005	0.008	0.006	0.007	0.006
Mg	0.921	0.867	1.059	1.058	1.025	1.141	0.938	0.913	0.883	0.908
Ca	0.821	0.856	0.765	0.820	0.800	0.714	0.832	0.804	0.821	0.801
Na	0.019	0.029	0.030	0.029	0.019	0.021	0.014	0.034	0.033	0.034
K	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.001	0.000
Wo	42.07	44.38	37.81	39.39	39.74	34.96	40.72	41.90	42.61	41.46
En	47.18	44.96	52.37	50.83	50.93	55.84	45.92	47.61	45.85	46.99
Fs	9.76	9.14	8.33	8.37	8.40	8.15	12.69	8.74	9.83	9.78
Mg#	82.95	83.51	86.23	85.88	86.22	87.47	78.64	84.96	82.86	83.21

Figure 3. Back-scattered (BSE) images of Pulivendla ultramafic rock showing olivine (Ol), pyroxene (Cpx), chrome-spinel (Cr-spl) substantiated with X-ray elemental map for variation of Ca (bottom right) and Al (bottom left).

equilibrium, effectively facilitate to estimate the crystallization temperature precisely (Wan *et al.* 2008; Coogan *et al.* 2014). Furthermore, it is less susceptible to resetting than Mg–Fe thermometry due to slow diffusion of Al^{3+} than Mg and Fe in olivine lattice (Spandler and O'Neill 2010). As olivine and chrome-spinel being the earliest phases to crystallize from primitive mantle melts, the Al-in-olivine thermometry provides an opportunity for precise estimation of peak temperature of co-precipitation of these phases; however, one should be very cautious to use this temperature as it is unlikely to match the liquidus temperature exactly (Coogan *et al.* 2014). Therefore, the maximum crystallization temperature of mantle derived melts provides constraints on the minimum temperature of their mantle source regions (Coogan *et al.* 2014). By using Al-in-olivine and chrome-spinel pair, that is in equilibrium, Wan *et al.* (2008) experimentally calibrated this thermometry at 0.1 MPa pressure and 1250–1450°C temperature with chrome-spinel compositions (Cr#) in the range of 0.07–0.69. Coogan *et al.* (2014) have further modified it by expanding calibration range (higher $f\text{O}_2$, spinel $\text{Fe}^{3+}/\text{Fe}_T \leq 0.35$). The Al-in-olivine thermometry $\{T(\text{K}) = 10^4/[0.575 + 0.884\text{Cr}^{\text{sp}} - 0.897 \ln(\text{Al}_2\text{O}_3^{\text{ol}}/\text{Al}_2\text{O}_3^{\text{sp}})]\}$ proposed by Coogan *et al.* (2014) has been used for the present study where the error is estimated to be within $\pm 25^\circ\text{C}$.

Olivine-spinel pair, satisfying boundary condition of this thermometry, is used for calculating crystallization temperature; which ranges from 1410° to 1484°C with an average of 1453°C. This average temperature is 260°C higher than the average temperature (1193°C) of MORB; even it is 183°C higher than the maximum (1270°C) temperature observed for MORB.

4.3 Estimation of mantle potential temperature

The mantle potential temperature (T_p) is believed to be the balance between heat loss by convective and conductive cooling through the earth's lithosphere and heat gain by radioactive heating in the mantle and conductive heating at the core-mantle boundary (Ganne and Feng 2017) and corresponds to a temperature of mantle material, which is adiabatically ascended to the surface without any melting (McKenzie and Bickle 1988). The T_p can be estimated using a wide range of techniques, such as interpreting REE inversion, whole-rock compositions and melt inclusions (McKenzie and O'Nions 1991; Herzberg and Asimow 2015) and can be

compared with the crystallization temperature for better understanding towards magma genesis. Anand *et al.* (2003) have assessed T_p of the Pulivendla ultramafic–mafic sills using REE inversion model, which estimates a temperature of $\sim 1500^\circ\text{C}$. In the present study, T_p is estimated using major element data by a computer program PRIMELT3 (Herzberg and Asimow 2015) for the studied Pulivendla sills (Anand *et al.* 2003). This could help to model batch as well as accumulated fractional melting by adjusting olivine to match with a peridotite melt composition (Herzberg and Asimow 2015). The ultramafic samples, having $\text{MgO} > 24$ wt% have been considered for PRIMELT3 calculations. There is no significant variation observed in the calculated melt composition, eruptive temperature and T_p under reducing ($\text{Fe}_2\text{O}_3/\text{TiO}_2 = 0.5$) and oxidizing ($\text{Fe}_2\text{O}_3/\text{TiO}_2 = 1$) conditions. For the present case, Fe^{2+}/Fe ratio in the mantle peridotite was kept at 0.9. Obtained results suggest that primary melt is komatiitic in nature ($\text{MgO} > 20$ wt%; average 26.4 wt%) and equilibrium with olivine containing $\text{Fo}_{92.2}$. The average T_p is 1688°C with an uncertainty of $\pm 60^\circ\text{C}$ (2σ) (Herzberg and Asimow 2015).

Besides ~ 1.88 – 1.89 Ga Hampi–Bastanar LIP, which also includes the Pulivendla sills, seven other mafic dyke swarms, thought to be associated with distinct LIP events, have been identified in the Dharwar craton (cf. Samal *et al.* 2019; Söderlund *et al.* 2019). An adequate amount of geochemical data is available for the ~ 2.37 Ga Bangalore–Karimnagar, ~ 2.22 Ga Kandlamadugu, and ~ 2.21 Ga Anantapur–Kunigal swarms (French and Heaman 2010; Srivastava *et al.* 2014a, b, 2015). For comparison, T_p is also estimated using the available geochemical data for these swarms. Samples with high MgO (wt%) content and Mg# [EDD 09/02, 2.37 Ga swarm (Srivastava *et al.* 2014a), EDD 09/04, 2.22 Ga swarm (Srivastava *et al.* 2014b) and JEF 99/11, 2.21 Ga swarm (French and Heaman 2010)] are considered for estimating T_p . Mantle potential temperature of 1765°, 1628° and 1564°C, respectively, is estimated for these swarms, which are significantly high and almost analogous to the estimation for the Pulivendla sills of the present study.

4.4 Nature of parental magma

The composition of early formed phases (viz., chrome-spinel and olivine) also reflects the melt composition from which they were crystallized

(Dick and Bullen 1984; Barnes and Roeder 2001; Kamenetsky *et al.* 2001). The linear relationship between Al_2O_3 and TiO_2 content of chrome-spinel and the respective Al_2O_3 and TiO_2 concentrations in the melt (Maurel and Maurel 1982; Kamenetsky *et al.* 2001) is the fundamental principle for calculating the parental melt composition. Based on this idea, Al_2O_3 and TiO_2 content in the parental melt of the Pulivendla ultramafic–mafic sills have been estimated (see table 2). The equation Al_2O_3 wt% in chrome-spinel = $0.035 \times \text{Al}_2\text{O}_3^{2.42}$ in parental melt, proposed by Maurel and Maurel (1982) for the spinel-liquid equilibrium at 1kbar, is used to calculate Al_2O_3 content of the parental melt. Results suggest that Al_2O_3 contents of the parental melt of the ultramafics of the Pulivendla sills range between 8.43 and 11.76 wt%, which is well-comparable to parental melts of other occurrences, such as the Bushveld and the Great Dyke of Zimbabwe (Wilson 1982; Sharpe and Hulbert 1985). Similarly, TiO_2 content of the parental melt is calculated by using the equation TiO_2 in parental melt = $1.0963 \times \text{TiO}_2^{0.7863}$ in chrome spinel (Rollinson 2008). The parental melt composition of Pulivendla sills is significantly rich in TiO_2 contents, ranging from 1.38 to 5.42 wt%.

5. Discussion

5.1 Role of subcontinental lithospheric mantle (SCLM)

5.1.1 Mineral chemistry

Composition of chrome-spinel can be used as an important petrogenetic indicator, particularly for high-Mg rich mafic rocks, which may help in discriminating their tectonic setting (Kamenetsky *et al.* 2001). Barnes and Roeder (2001), based on global chrome-spinel database, have defined compositional fields for a wide range of mafic–ultramafic rocks derived from variable tectonic settings. As the studied chrome-spinel contains moderate amount of aluminum and iron, they are classified as Al- to Fe-chromite in Cr–Al–Fe³⁺ ternary discrimination diagram (figure 4a; after Stevens 1944). The high Ti (0.320–1.625) and Fe³⁺ (2.388–4.429) coupled with relatively low Al (1.856–4.270) apfu suggests distinctive characteristics of Alaskan-type spinels (see figure 4b; Kepezhinskis *et al.* 1993). It also has intermediate Cr# (av. ~0.71), relatively low Mg# (av. ~0.29) and high Fe³⁺# (av. 0.23), which suggest its

Alaskan-type nature and that can be well distinguished from ophiolitic and other settings of ultramafic rocks (see figure 5a and b). The abundance of Al_2O_3 and TiO_2 in magmatic spinel is mainly controlled by contents of these oxides in the parental melt, hence can be used to discriminate their tectonic affinities (Kamenetsky *et al.* 2001). The high TiO_2 content in the parental melt of the studied rock suggests that it was initially Ti-rich; well supported by high TiO_2 content in the early forming phase spinel. On TiO_2 vs. Al_2O_3 diagram (figure 5c; after Kamenetsky *et al.* 2001), studied spinels plot in both LIP and OIB field and many overlap with Alaskan-type. Rollinson (2008) has defined relationships between melt– Al_2O_3 and spinel– Al_2O_3 for MORB and arc lavas; Power law expression for MORB lava and logarithmic expression for arc lava. Spinel of the present study are plotted significantly away from the MORB trend and show arc type characteristics (not shown). NiO and MnO contents of olivine are in the range of global Alaskan-type olivine (see figure 6a, b). The higher Al (av. 149 ppm), Ca (av. 918) and Ti (av. 152 ppm) content correspond to the typical characteristics of igneous olivine as lower values of Al (<130 ppm) and Ca (<700 ppm) suggestive of mantle olivine (figure 6c, Foley *et al.* 2013). Furthermore, trace element variations suggest that the olivine grains are not entrained xenocryst and crystallized from the parental magma. A comparative chart of Fo-content of typical Alaskan-type ultramafic rocks is given in figure 6(d). The Fo-content of the studied Pulivendla ultramafic rocks is well-comparable with those from typical Alaskan-type complexes. The relationship between the Cr# of spinels and Fo-contents of olivine is defined by the olivine-spinel mantle array (OSMA), which also accommodates residual and refractory peridotites (Arai 1994). Peridotites of cumulate origin generally plot in low-Fo area, away from the OSMA. Pulivendla ultramafic samples plot in Alaskan-type field and follow the cumulate trend (figure 7). The Ca-rich nature and variation of Al_2O_3 content in clinopyroxenes of the studied ultramafic samples (figure 8a, d) are also comparable with the Alaskan-type complexes (Farahat and Helmy 2006; Helmy *et al.* 2014, 2015). On the discrimination diagram based on Al_2O_3 and TiO_2 (Loucks 1990) and Al_2O_3 and SiO_2 content (Le Bas 1962), the studied clinopyroxenes display arc-cumulated trend of Alaskan-type intrusions (figure 8b, c). Therefore, mineral chemistry, in general, display

Figure 4. (a) Cr–Al–Fe³⁺ (after Stevens 1944) and (b) Al–Cr–Fe³⁺+Ti ternary plots for spinels from various tectonic settings. Compositional fields for spinels from ophiolites and Alaskan-type complexes are from Jan and Windley (1990).

typical characteristics of an Alaskan type or arc type of melt.

5.1.2 Whole-rock geochemistry

Generally, Alaskan-type complexes are characterized by arc magma chambers (Taylor 1967; Irvine 1974) or arc-root complexes (DeBari and Coleman 1989) and show distinguished morphological, petrographical, mineralogical and geochemical characteristics (Habtoor *et al.* 2016). Chemical diversity among Alaskan-type complexes can be attributed to the variable inputs of subduction component across the arc (Fershtater *et al.* 1997). Therefore, the observed mineralogical characteristics, as discussed above, could be due to the involvement of a subducted component in the mantle source. To delineate this inference, available geochemical data (e.g., Anand *et al.* 2003) of the Pulivendla ultramafic–mafic rocks are discussed to characterize the nature of the mantle source. Some key trace element ratios (HFSE/LREE) are significant indicator to categorize the mantle source region (Smith *et al.* 1999). The lower Nb/La ratio (< 0.3) of the studied rocks is consistent with the nature of a lithospheric mantle (Nb/La < 0.5) rather than an OIB like asthenospheric mantle (Nb/La > 1) (figure 9). The studied samples show enriched primitive mantle normalized large-ion lithophile elements (LILE) and LREE and display negative Nb–Ta and Ti anomaly on multi-element spidergrams similar to different Paleoproterozoic mafic dyke swarms of the Dharwar craton (see figure 10). This significant enrichment can be attributed due to the involvement of a crustal component en route to the surface or source enrichment through subduction zone processes

(Saunders *et al.* 1980, 1991; Sun *et al.* 1989; Zhao and McCulloch 1993; Zhao and Asimow 2018). The studied samples also exhibit a large range of Nb/La with limited variations in Nb/Th ratios, which contradicts the magma contamination model as addition of crustal material would reduce both the Nb/La and Nb/Th ratios (Zhang *et al.* 2011). Also the studied samples show high Ba/Nb, Ba/Zr, Ba/Th and Th/U ratios relative to the primitive mantle (Sun and McDonough 1989) and as well as the average upper crust (Rudnick and Fountain 1995). These observations negate any significant role of crustal contamination and probably suggest their derivation from a lithospheric mantle modified through an ancient subduction event. This inference is also well-supported through geochemical and isotopic studies on the Mesoproterozoic lamprophyre dykes, exposed to the western side of the Cuddapah basin (Pandey *et al.* 2017a, b, 2018).

The addition of subducted components into overlying mantle wedges can be possible either through mobile fluids (Hawkesworth *et al.* 1997; Turner *et al.* 1997; Class *et al.* 2000) or melts (Stern and Kilian 1996; Elliott *et al.* 1997; Hawkesworth *et al.* 1997; Münker 2000; Stracke *et al.* 2003). The enrichment of relatively immobile REE and HFSE (e.g., Th, Zr, Hf, Nb and Ta) in a mantle source is controlled by the introduction of slab melts (Tatsumi 1989; Keppler 1996; Turner *et al.* 1997), whereas large ion lithophile elements (LILE, e.g., Rb, Ba, U) and H₂O in arc magmas is controlled by fluid influx (Stolper and Newman 1994). Therefore, ratios of LILE/REE and LILE/HFSE provide significant input to study role of fluids and melts in the petrogenesis of a given arc magmas. The mantle sources modified by slab melts are likely to have lower Th/Zr, Rb/Y, Ba/Nb and Ba/

Figure 5. (a) Mg# vs. Cr# and (b) Mg# vs. Fe³⁺ diagrams for studied chrome-spinels. Different compositional fields are taken from boninite (Barnes and Roeder 2001); abyssal peridotite (Dick and Bullen 1984); Alaskan-type field (Burns 1985, Himmelberg and Loney 1995); Alpine-type, stratiform and Alaskan-type complexes (Irvine 1967). (c) TiO₂ (wt%) vs. Al₂O₃ (wt%) discrimination diagram of spinels derived from different tectonic settings (after Kamenetsky *et al.* 2001). Compositional fields of Alaskan-type complexes are from Himmelberg and Loney (1995). Himmelberg *et al.* (1986), basalts of large igneous provinces (LIP), ocean island basalts (OIB), mid-ocean ridge basalts (MORB), and island-arc series are from Kamenetsky *et al.* (2001).

Th ratios than those dominated by fluid input (Hawkesworth *et al.* 1997; Kepezhinskas *et al.* 1997). On Th/Zr and Nb/Zr variation plot, the studied samples follow the trend of slab fluid induced metasomatism (figure 11a). Similar inference is also observed on Nb/Y and Rb/Y ratio plot (figure 11b). These results suggest that the lithospheric mantle might have been influenced by subduction-related fluid. Moreover, the presence of

primary phlogopite in the studied rock also complements these observations.

Similarly, Ba/La together with La/Sm ratios are considered to be the indicator of slab fluid inputs in the subduction zones (Ryan *et al.* 1996); these ratios in studied mafic and ultramafic samples suggest subduction-induced fluid enrichment, rather than crustal contamination (figure 12a). On the AFM diagram (Beard 1986), studied samples

Figure 6. (a) Variation of NiO (wt%) against Fo (mole %) content in olivine. Data sources: Alaskan global trend (Krause *et al.* 2007); Gabbro Akarem (Helmy and Mogessie 2001; Helmy and El Mahalawi 2003); Dahaniib (Khedr and Arai 2016). (b) Variation of MnO (wt%) with Fo (mol%) content in olivine. Data sources: Genina Gharbia (Helmy *et al.* 2008, 2014); Gabbro Akarem (Helmy and Mogessie 2001; Helmy and El Mahalawi 2003); Dahaniib (Khedr and Arai 2016). (c) Variations of Ca (ppm) and Ti (ppm) in olivine; Mantle and Igneous olivine values are from Foley *et al.* (2013). (d) Forsterite (Fo) content of olivines of present study compared with the typical Alaskan-type complexes reported worldwide (modified after Pettigrew and Hattori 2006). Data sources: Quetico Intrusions (MacTavish 1999), Turnagain (Clark 1980), Gabbro Akarem Complex (Helmy and Mogessie 2001), Blashke Island (Himmelberg *et al.* 1986), Union Bay, Polaris, and Duke Island (Irvine 1974, 1976), the Tulameen Complex (Rublee 1994), the Taftafan Complex (Habtoor *et al.* 2016), Alaskan Complexes (Irvine 1974).

also show arc-related cumulate ultramafic signature (see figure 12b). Similar geochemical characteristics are also noticed in different generation of Paleoproterozoic mafic dyke swarms (figure 10) including ~ 1.88 – 1.89 Ga Hampi-Bastanar swarm (French *et al.* 2008; Samal *et al.* 2019) spread over the Dharwar (e.g., Srivastava *et al.* 2014a, b, 2015; Söderlund *et al.* 2019) as well as the Bastar cratons (e.g., Srivastava and Gautam 2015). It is suggested that observed geochemical characteristics can be attributed to partial melting of a SCLM that has previously been modified by subduction related fluids (e.g., Sandeman and Ryan 2008; Ernst 2014).

5.2 Decoding Alaskan-type signature of the Pulivendla ultramafic-mafic sills

Alaskan-type complexes signify a group of ultramafic bodies with a concentrically zoned distribution, evolving from a dunite core to a

clinopyroxenite rim, with gabbros or diorites at the margins and usually associated with platinum-group elements (PGE) mineralization (Taylor 1967; Wyllie 1967; Taylor and Noble 1969; Foley *et al.* 1997). The size of Alaskan-type complexes ranges from a few to hundreds of km and in map view and show circular or elliptical geometries (Guillou-Frottier *et al.* 2014). They can appear as isolated bodies or accompanied by a twin body of similar size, each of them being separated by a few kilometers and excludes large intrusions (Guillou-Frottier *et al.* 2014). Its association is observed both in ancient subduction zones (Urals, Alaska, British Columbia, Kamchatka, Colombia) and in intracontinental settings (Siberia, Ethiopia, Far East Russia), however chemical (mineralogical and geochemical) diversity among Alaskan-type complexes can be attributed to the variable inputs of subduction component (Fershtater *et al.* 1997). Therefore, it also acts as a tracer of an ancient subduction event in earth's history (Su *et al.* 2012).

Figure 7. Variation of Mg# of olivine and Cr# of spinel. The olivine-spinel mantle array (OSMA) is shown by the two bold lines with a supra-subduction zone (SSZ), abyssal peridotites and passive margin peridotites fields (Dick and Bullen 1984; Dick 1989; Arai 1994; Pearce *et al.* 2000). Alaskan-type field from Blades *et al.* (2019).

Further, as discussed above, the studied Pulivendla ultramafic–mafic sills is a part of a widespread ~1.88–1.89 Ga Hampi–Bastanar LIP event (cf. Samal *et al.* 2019) covering an area of about 90,000 km², which is not straightforward to support an Alaskan-type complex. Also, its geometry, morphology and modal mineralogy (presence of orthopyroxene) is not supportive to Alaskan-type of intrusion (see table 5 for detailed comparison). Therefore, the observed mineralogical and geochemical characteristics of the Pulivendla ultramafic–mafic rocks, showing Alaskan-type ultramafic–mafic intrusions, can only be due to a metasomatized mantle source through an ancient subduction event.

Existence of an ancient subduction event in the Dharwar craton is established through several geochemical and isotopic studies on felsic and mafic volcanic rocks from the Neoarchean greenstone belts (e.g., Manikyamba *et al.* 2005; Ram Mohan *et al.* 2013; Khanna *et al.* 2015, 2016). It is suggested that these Neoarchean felsic (adakitic-affinities and Mg-andesites) and mafic (boninite-like ultramafics) volcanic rocks have their genesis through subduction tectonics. Based on available field, petrographic, geochronologic, and elemental and Nd–Sr–Pb–Hf isotope data for TTG gneisses, transitional TTGs, sanukitoids and anatectic granites from the Dharwar Craton, Jayananda

et al. (2018) have reviewed the accretionary processes of juvenile crust, continental growth and evolving geodynamic processes through the 3.45–2.5 Ga. They have suggested that the melting of slabs with detritus at great depth during ca. 2.7 Ga generated magmas, which upon their upward passage enriched arc mantle. Therefore, the Neoarchean subduction signature in the studied ca. 1.89 Ga Pulivendla rocks is consistent with the previously proposed uniformitarian models for the evolution of EDC.

5.3 Evidence of ca. 1.88–1.89 Ga LIP and associated mantle plume in the Dharwar Craton (and Cuddapah basin)

The nature of the mantle source and its subsequent melting are two independent processes and difficult to assess the time gap between these two, therefore, subduction-modified lithospheric mantle can provide a source for subsequent magmatism either immediately after subduction or even after a prolonged time gap (Johnson *et al.* 1978; Hall and Hughes 1993; Goodenough *et al.* 2002). It is suggested that mafic magmatic style have been changed during the Archean, which is evident from waning phase of the komatiite magmatism towards the end of the Neoarchean followed by a widespread occurrence of high-Mg mafic magmatism (Hall and Hughes 1993). Most importantly, this change was due to crustal thickening and growth and stabilization of the cratonic lithosphere (cf. Hall and Hughes 1993; Srivastava *et al.* 2016).

Most of the LIPs in geological record are supposed to be associated with plume driven magmatism in an intracratonic setting. It is difficult to explain generation of a large volume of magma to feed an igneous province by any other mechanism (cf. Ernst 2014). Because of this very reason, many have favoured plume tectonics for the genesis of the studied Pulivendla sill, a part of the ca. 1.88–1.89 Ga Hampi–Bastanar LIP and emplaced within the Cuddapah basin (French *et al.* 2008; Ernst and Srivastava 2008; Shellnut *et al.* 2018; Samal *et al.* 2019; Srivastava and Samal 2019; Söderlund *et al.* 2019). Therefore, it is believed that impingement of a mantle plume could be the causative factor for the partial melting of a SCLM modified during an ancient subduction event below the eastern Dharwar Craton and Cuddapah basin at ca. 1.88–1.89 Ga (figure 13).

Figure 8. (a) Classificatory triangular diagram of Cpx (after Morimoto 1989). (b) Al_z ($Al^{IV} \times 100/2$) vs. TiO_2 (wt%) of Cpx compositions after Loucks (1990). The fields of Gabbro Akarem (Helmy and El Mahallawi 2003), Quetico Intrusions (Pettigrew and Hattori 2006), Alaskan type Intrusions (Himmelberg and Loney 1995) are also plotted for comparison. (c) SiO_2 (wt%) vs. Al_2O_3 (wt%) discrimination diagram (after Le Bas 1962). (d) Al_2O_3 (wt%) vs. Mg# of Cpx. Data for Alaskan-type intrusions is taken from Himmelberg and Loney (1995).

Figure 9. Variation of La/Yb vs. Nb/La to illustrate the role of lithosphere/asthenosphere in the gneiss of the studied ultramafic-mafic rocks (after Smith *et al.* 1999).

There are several criteria to identify the existence of a mantle plume (Ernst and Buchan 2003). The aerial extent, available precise age

data, and radiating pattern of dyke swarm obviously support the presence of ca. 1.88–1.89 Ga Hampi–Bastanar LIP (which also includes Pulivendla sills) in the Indian shield and its genesis through plume-induced magmatism (French *et al.* 2008; Ernst and Srivastava 2008; Shellnut *et al.* 2018; Samal *et al.* 2019; Srivastava and Samal 2019; Söderlund *et al.* 2019). Recently, the newly identified ca. 1.89 Ga Boonadgin dyke swarm of the Yilgarn Craton (Stark *et al.* 2019) has also been correlated with the Hampi–Bastanar LIP (Samal *et al.* 2019). Based on paleomagnetic results, Liu *et al.* (2019) have suggested that the Boonadgin dyke swarm of the Yilgarn Craton represent an arm of a radiating dyke swarm that shared the same plume centre with coeval mafic dykes in the Dharwar and Bastar Cratons, and Pulivendla sill of the Cuddapah basin.

Similarly, magmatic underplating is the deepest and most concealed component of a plumbing system in which mantle-derived magma accumulates at

Figure 10. (a) Primitive mantle normalized multi-element spidergram, (b) chondrite normalized rare earth element pattern is the average of all the available geochemical data. Data sources: 1.89 Ga Pulivendla sill (Aannd *et al.* 2003); 2.37 Ga Bangalore–Karimnagar swarm (Srivastava *et al.* 2014a); 2.21 Ga Anantapur–Kunigal swarm (Kumar *et al.* 2012a, b; Srivastava *et al.* 2014b); 2.18 Ga Mahbubnagar–Dandeli swarm (Srivastava *et al.* 2015); 1.88–1.89 Ga Hampi swarm (Srivastava *et al.* 2014a).

the base of the crust (Cox 1993). This may have relation with seismically high velocity lower crust (HVLC) and is generally associated with LIPs (Ernst 2014). The formation of a lopolitic or a funnel-shaped magma cupola in the upper crust along southwestern part of the Cuddapah basin has been observed through high gravity anomaly which may be the manifestation of an asthenospheric upwelling (Chatterji and Bhattacharji 2001). Such domal uplift is one of the most prominent features to recognize the existence of a mantle plume (Ernst and Buchan 2003). Also, based on 2-D modelling of seismic structure together with other geological and geophysical signatures across the Cuddapah basin, Mall *et al.* (2008) have suggested the presence of a mantle plume during Proterozoic. High gravity and conductivity anomaly along with 15–20 km thick magmatic underplating that results the thickening of the crust to about 40–44 km below these regions

provides compelling evidence for a deep seated mantle plume below southwestern part of the Cuddapah Basin (Chandrakala *et al.* 2010).

Komatiite (MgO >18%) and picrite (MgO ranges between 12 and 18 wt%) are important components of any LIP because, in general, they are generated from the hottest portion of a plume immediately above plume conduit (Campbell *et al.* 1989). The Proterozoic threshold for identifying plume related magma has been set at 12 wt% (Gibson 2002). The studied rocks have MgO content >20 wt% that strongly supports plume signature.

5.3.1 Evidence from crystallization and mantle potential temperatures

Defining thermal regime of the mantle may provide significant evidence for the existence of a mantle plume as it is characterized by thermal anomalies in

Figure 11. Variation of (a) Th/Zr vs. Nb/Zr and (b) Nb/Y vs. Rb/Y for the studied ultramafic-mafic rocks to evaluate the role of fluid/melt induced metasomatism.

the upper mantle, which is supposed to be 260°C higher than the ambient mantle (1300°–1400°C) (Farnetani and Richards 1994; Herzberg and O'Hara 2002; Herzberg *et al.* 2007). Coogan *et al.* (2014) have studied olivine-spinel co-saturation temperature from four LIPs around the globe which provides an average crystallization temperature of 1420°C. Similarly, Heinonen *et al.* (2015) have suggested a crystallization temperature between 1391±42°C and 1481±35°C, based on olivine-spinel thermometry, for Karoo LIP. Also, Xu and Liu (2016) have estimated a maximum crystallization temperature of 1440±63°C for Emeishan LIP. The estimated maximum crystallization temperature of 1484°C for the studied rock is significantly higher than those of MORB and also analogous to the crystallization temperature of these LIPs (see figure 14a) and suggests that a mantle source was heated above the ambient upper mantle temperature.

Figure 12. (a) Variation of La/Sm vs Ba/La (after Ryan *et al.* 1996) to infer the effect of en route crustal contamination or fluid/melt related metasomatism in the source. Data sources: Upper Crust (UC; Taylor and McLennan 1985; Rudnick and Gao 2003), Lower Crust (LC; Weaver and Tarney 1984); (b) AFM diagram for the Pulivendla ultramafic rocks. Different tectonic fields for cumulate and non-cumulate mafic-ultramafic rocks are taken from Beard (1986).

The conversion of crystallization temperature into mantle potential temperature is based on several assumptions and requires corrections on depth of melting and latent heat of fusion of the mantle (e.g., Green and Falloon 2005; Falloon *et al.* 2007a, b; Putirka *et al.* 2007). Matthews *et al.* (2016) have developed a thermal model of mantle melting to quantify uncertainties in converting crystallization temperature to mantle potential temperature. It suggests that a crystallization temperature of 1399±20°C for the Iceland volcanic rocks corresponds to a mantle potential temperature of ~1500°C. Similarly, Al-in-olivine crystallization temperature of 1440°C for Emeishan LIP (Xu and Liu 2016) corresponds to a mantle potential temperature >1500°C (Shellnutt and Pham 2018). Therefore, the estimated maximum crystallization temperature 1484°C for the studied ultramafic rocks can be extrapolated to a mantle potential

Table 5. Comparison between worldwide Alaskan-type intrusions (Pettigrew and Hattori 2006) with studied Pulivendla ultramafic–mafic sills.

Characteristics	Alaskan type intrusions	Pulivendla ultramafic-mafic rock (this study)
Age	Mostly Phanerozoic	~1.89 Ga
Distribution	Clusters of intrusions form long linear arrays along major tectonic boundaries in orogenic belts	Sills
Geological setting	Close to the end of subduction, prior to accretion-collision	In an intraplate tectonic setting, after the subduction event
Size	Most are small in size (<4 km)	90,000 km ²
Morphology and zoning	Crude concentric zoning of lithologies grading from olivine-rich ultramafic cores to mafic rims, lack of chilled margins	No lithological zoning
Sequence of intrusion	Late phase intrusion of gabbroic and dioritic rocks	No dioritic intrusion
Lithology	Dunite, hornblende and clinopyroxenite, gabbro and minor dioritic and syenitic rocks	Peridotite and Gabbro
Textures	Poikilitic textured hornblende in clinopyroxenite, common occurrence of mafic pegmatites	Poikilitic texture
Mineralogy	Abundant clinopyroxene, hornblende. Common occurrence of secondary magnetite, Essential lack of orthopyroxene	Olivine, occurrence of both Clinopyroxene and orthopyroxene
Presence of Chromite	Common occurrence of chromite in dunite	Occurrence of Chromite
Mineral chemistry	High Mg olivine, diopsidic clinopyroxene, phlogopitic mica. Amphibole is calcic showing, a wide range in composition	No high Mg Olivine, diopsidic clinopyroxene
Bulk rock geochemistry	High LILE and low HFSE Fractionated REE with no Eu anomalies	High LILE and low HFSE Fractionated REE with no Eu anomalies
PGE mineralization	PGE mineralization in olivine-rich cores (dunite) associated with chromite or Cu–Ni sulphides	No PGE mineralization reported yet

Figure 13. Cartoon showing emplacement of Pulivendla sills by melting of a modified sub-continental lithospheric mantle due to ca 1.88–1.89 Ga mantle plume (not to scale).

temperature of ~1600°C (figure 14b). However, for better understanding and correlation, a mantle potential temperature of 1688°C is also estimated using the geochemical data of Anand *et al.* (2003) for the studied rocks. This temperature is comparable

(within the error range) with the T_p extrapolated from the crystallization temperature (~1600°C) (figure 14b). Furthermore, the estimation of mantle potential temperature (T_p) and its comparison with the secular cooling model of the earth at 1.89 Ga also substantiates this assumption. This estimated T_p has been compared with available secular cooling models (figure 15; Richter 1988; Abbott *et al.* 1994; Davis 2009; Herzberg *et al.* 2010) and it records a higher estimate than all these models. Further, this is also valid for a recent model given by Herzberg *et al.* (2010) that suggests ~300°C hotter mantle than other models at 1.89 Ga (figure 15).

All these models together with calculated T_p for the Pulivendla sills infer that melt for these sills has been generated from a higher thermal regime in the upper mantle at 1.89 Ga, which can only be explained with a mantle plume. Further, this estimated T_p is also consistent with the T_p of ca 2.37 Ga Bangalore–Karimnagar, ca 2.22 Ga Kandlamadugu, and ca 2.21 Ga Anantapur–Kunigal swarms emplaced within the Dharwar Craton, which are supposed to be part of different LIPs (see figure 15).

Figure 14. (a) Histogram showing correlation of Al-in-olivine crystallization temperatures for MORB and LIPs. Data source: Karoo LIP (Heinonen *et al.* 2015), Baffin Island, SE Greenland, Madagascar, Gorgona Island (Coogan *et al.* 2014), Emeishan LIP (Xu and Liu 2016), MORB (Coogan *et al.* 2014). (b) Histogram showing comparison of crystallization and mantle potential temperatures (extrapolated). Data source: Emeishan LIP (Xu and Liu 2016; Shellnutt and Pham 2018), Island Rocks (Matthews *et al.* 2016).

Figure 15. Various models for thermal evolution of upper mantle through time (Modified after Ciborowski *et al.* 2017) compiling models given by (1) Davies (2009); (2) Richter (1988); (3) Abbott *et al.* (1994); (4) Herzberg *et al.* (2010). Average T_p of ~1.88–1.89 Ga Pulivendla sills plotted together with ~2.37 Ga Bangalore–Karimnagar, ~2.22 Ga Kandlamadugu and ~2.21 Ma Anantapur–Kunigal swarms for comparison.

6. Conclusions

- Mineral chemistry of early formed minerals, such as olivine, chrome-spinel and clinopyroxene of Pulivendla ultramafic cumulate rocks, intruded within the Cuddapah basin, suggest Alaskan-type characteristics.
- Higher Ba/Nb and low Nb/La point to their derivation from an enriched lithospheric mantle source and higher Th/Yb ratio along with negative Ti–Nb–Ta anomalies on the primitive mantle-normalized multi-element spidergrams indicate dehydrated fluids from the subducting slab could be the possible metasomatic agent.
- Although geochemistry and mineral chemistry of these rocks suggest subduction signature, their unambiguous association with ca. 1.88–1.89 Ga Hampi–Bastanar LIP provides testimony for plume related magmatism. The latter inference is further substantiated by the estimated crystallization temperature (1410°–1484°C) and mantle potential temperature (>1600°C).
- The widespread ca 1.88–1.89 Ga mafic magmatism in the Indian shield, well defined radiating geometry of associated mafic dyke swarms and available geophysical evidences (domal uplift, magmatic underplating) also favour for mantle plume related magmatism.

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