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1 **Origin of the oldest (3600-3200 Ma) cratonic core in the Western Dharwar Craton,**
2 **Southern India: Implications for evolving tectonics of the Archean Earth**
3

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88 *Abstract*

89 This contribution presents a comprehensive synthesis on the origin of oldest (3600-3200 Ma)
90 cratonic core in the Western Dharwar Craton (WDC), Southern India based our new field,
91 geochronologic, elemental and Nd-Hf isotopes data on the Holenarsipur greenstone belt and
92 adjacent granitoids combined with published record on the stratigraphically equivalent
93 greenstone belts in the same region. Our study shows that the oldest cratonic core in the WDC
94 formed through assembly of different tectonic units close to 3200 Ma. These tectonic units
95 include microcontinental remnants with oceanic plateaus, oceanic island arcs and preserved
96 oceanic crust similar to modern oceanic crustal section close to a spreading center. Isotopic age
97 data for these greenstone sequences (detrital zircons -3600-3230 Ma; volcanics 3384-3200 Ma)
98 and adjoining granitoids (3430-3400 and 3350-3270 Ma with remnants of older 3600-3500 Ma
99 gneisses) indicate that a large part of the old TTG-type granitoids in the WDC is either coeval
100 to or slightly younger than associated greenstone units. The final stage of assembly of these
101 tectonic elements into cratonic framework through horizontal motion of intervening oceanic
102 crust and eventual slab breakoff is marked by the formation of ca. 3200 Ma trondhjemite
103 magmas, emplacement of which into the lower crust caused partial convective overturn of the
104 crust, thereby leading to the development of dome and keel patterns followed by
105 metamorphism and cratonization of the WDC crust at 3100 Ma. Our study strongly suggests
106 that the tectonic environments in which Archean cratons formed require some sort of horizontal
107 motion but not necessarily modern plate tectonics. However, vertical addition of juvenile crust
108 in hotspot environments associated with mantle plumes played a major role in building of early
109 crustal nuclei. The model proposed for the formation of the cratonic core in the western

110 Dharwar craton is compatible with observations made globally in most of the Paleoproterozoic
111 cratons.

112 **Key words:** Archean tectonics; Dharwar Craton; TTGs; Komatiites; U-Pb zircon ages; Sm-Nd
113 and Lu-Hf isotopes; Whole rock geochemistry.

114

115 **1. Introduction**

116 The origin of Archean cratons is one of the major research themes in solid earth sciences as
117 they are known for great economic mineral potential as well as provide important insights into
118 the early earth dynamics, redox evolution of ocean-atmospheric system and emergence of
119 biosphere ([Hashizume et al., 2016](#); [Johnson et al., 2017](#); [Catling and Zahnle, 2020](#); [Windley et
120 al., 2021](#); [Bosak et al., 2021](#); [Korenaga, 2021](#)). Archean cratons differ from their present-day
121 counterparts by exhibiting strong dichotomous lithological assemblages made of greenstone
122 belts (ultramafic-mafic and felsic volcanic rocks together with detrital and chemical
123 sedimentary rocks) and adjacent granitoids (dominantly made of TTGs; tonalite-trondhjemite-
124 granodiorite, e.g., [Anhaeusser, 2014](#)). The tectonic context of formation of both lithological
125 assemblages and the processes responsible for their spatial association are still disputed (e.g.,
126 [Smithies et al., 2019](#); [Adams et al., 2012](#); [Moyen and Martin, 2012](#); [Kaczmarek et al., 2016](#);
127 [Bedard, 2018](#); [Wyman, 2018](#)). Several studies argue for horizontal motion of lithospheric plates
128 initiated between 3200 Ma and 2500 Ma (e.g., [Gerya, 2014](#); [Cawood et al., 2018](#); [Wiemer et
129 al., 2018](#)) whereas others presented evidence for craton specific Eoarchean horizontal tectonics
130 ([Polat et al., 1998, 2015](#); [Kusky and Polat, 1999](#); [Komiya et al., 2015](#); [Hastie and Fitton, 2019](#);
131 [Nutman et al., 2021](#); [Garde et al., 2020](#); [Roman and Arndt, 2020](#); [Korenaga, 2021](#); [Kusky et
132 al., 2021](#); [Windley et al., 2021](#); [Grocolas et al., 2022](#); [Sotiriou et al., 2022](#)). Alternative models
133 involve stagnant crustal lid with periodic mantle overturn resulting in mantle upwelling, in
134 turn, generating komatiitic to basaltic volcanism forming volcanic plateaus which were

135 subsequently remelted to form felsic crust with eventual delamination of garnet-rich residue
136 (e.g., [Bedard, 2018](#)). Recently proposed models on Archean tectonics involve stagnant lid,
137 crustal drips, and horizontal motion of stagnant lid ([Bedard, 2006, 2018](#); [Rozel et al., 2017](#);
138 [Moyen and Laurent, 2018](#); [Johnson et al., 2018](#); [Nebel et al., 2018](#)). Part of the uncertainty
139 regarding the tectonic setting of Archean continent formation relies on when plate tectonics
140 began on the planet Earth, which is currently an unsolved issue and subject of much discussion
141 (e.g., [Turner et al., 2020](#); [Korenaga, 2013, 2021](#); [Windley et al., 2021](#)). Addressing this issue
142 is crucial as it bears on the thermal and chemical evolution of the early earth, the formation of
143 habitable continents and the emergence of biosphere.

144 The western Dharwar craton is an ideal target for studying the evolving early earth
145 dynamics, continental growth and craton formation partly because it forms a wide time and
146 tectonic window with a preserved crustal record of 3450 – 2600 Ma ([Jayananda et al., 2018](#))
147 with remnants as old as of 3600 Ma and isotopic evidence for even older fractionation events
148 at 3800 Ma ([Meen et al., 1992](#); [Nutman et al., 1992](#); [Jayananda et al., 2015](#); [Guitreau et al.,](#)
149 [2017](#); [Ravindran et al., 2020](#)). The oldest (ca.3600-3300 Ma) preserved TTGs - greenstone
150 assemblages are located in the vicinity of Holenarsipur greenstone belt (HGB) correspond to a
151 cratonic core which was intruded by 3200 Ma diapiric trondhjemites (e.g., [Naqvi and Rogers,](#)
152 [1987](#); [Meen et al., 1992](#); [Jayananda et al., 2008, 2015, 2019](#); [Guitreau et al., 2017](#); [Ravindran et](#)
153 [al., 2020](#)) coinciding with the development of regional dome and keel structures ([Bouhallier et](#)
154 [al., 1993,1995](#); [Chardon et al., 1996](#)). Furthermore, 3600-3200 Ma detrital zircon record
155 together with 3400-3200 TTG- greenstone volcanic assemblages also preserved in the regions
156 adjoining to Holenarsipur greenstone belt ([see fig.2](#)) which include Nuggihalli, Kalyadi,
157 Jayachamaraja Pura (J.C. Pura), Banasandra, Nagamangala, southern margin of the Bababudan
158 basin and western margin of the Chitradurga greenstone belt ([Nutman et al., 1992](#);

159 Ramakrishnan et al., 1994; Hokada et al., 2013; Lancaster et al., 2015; Wang et al., 2020;
160 Jayananda et al., 2008, 2016, 2020; Ravindran et al., 2020).

161 However, the mechanisms and tectonic context(s) of crust formation and craton
162 building mechanisms in the oldest crustal nuclei of the WDC remain debated and have been
163 the subjects of much discussion (Kunugiza et al., 1996; Choukroune et al., 1997; Chadwick et
164 al., 2000; Jayananda et al., 2008, 2015; Naqvi et al., 2009; Guitreau et al., 2017; Patra et al.,
165 2020; Ranjan et al., 2020; Panicker et al., 2021). The greenstone sequences and the adjoining
166 granitoids of TTG affinity have been interpreted as oceanic plateau and island arc (Jayananda
167 et al., 2008, 2015, 2016; Peucat et al., 1995), flat subduction of plume-fed oceanic plateau crust
168 (Naqvi et al., 2009), or remnants of migrating oceanic crust from mid-ocean ridge to trench
169 (Kunugiza et al., 1996). The WDC is considered to be a collage of different stratigraphic units
170 wherein geological (Swami Nath and Ramakrishnan, 1981; Naqvi, 1983), strain fabrics data
171 (Bouhallier et al., 1993,1995; Chardon et al., 1996,1998,2008), petrological, elemental and
172 isotopic data have been published (Meen et al., 1992;Peucat et al., 1993, 1995; Naqvi et al.,
173 2009; Jayananda et al., 2008, 2013a, 2015; Guitreau et al., 2017; Dasgupta et al., 2019; Wang
174 and Santosh, 2019; Ranjan et al., 2020). The Holenarsipur greenstone belt and adjacent
175 granitoids (TTG-type Peninsular gneisses) have long been considered as early Precambrian
176 crustal nuclei (Radhakrishna and Naqvi., 1986). Komatiite-dominated volcanic sequences
177 stratigraphically equivalent to the Holenarsipur greenstone belt can be found in 3400-3200 Ma
178 Nuggihalli, Kalyadi, J.C. Pura, Ghattihosahalli, Banasandra and Nagamangala greenstone belts
179 (Chadwick et al., 1981; Swami Nath and Ramakrishnan, 1981; Venkata Dasu et al., 1991;
180 Jayananda et al. 2008, 2016, 2020). However, spatial link between greenstone volcanism and
181 adjoining granitoids formation, mantle evolution and continental growth, evolving tectonics
182 and craton building processes are still largely conjectural. No major multidisciplinary approach
183 has been initiated to address the origin of the oldest cratonic core in the WDC. Therefore, the

184 older greenstone sequences (Sargur Group) and their adjoining granitoids (TTGs) form a most
185 relevant target to address evolving early earth dynamics, mantle evolution, continental growth
186 and origins of cratonic cores worldwide.

187 Here, we present a comprehensive synthesis based on multidisciplinary approach that
188 involves our new field, petrological, and geochemical data (whole-rock elemental and Sm-Nd
189 isotope data as well as U-Pb zircon ages coupled with *in-situ* Lu-Hf isotope data) in
190 combination with published data for the Holenarsipur greenstone belt and adjacent granitoids
191 (TTG-type Peninsular gneisses) and diapiric trondhjemite plutons as well as published record
192 from the other stratigraphically equivalent Sargur Group greenstone belts and their spatially
193 associated granitoids of TTG-affinity forming cratonic core in the WDC (see [fig.1 and 2a](#)).
194 We address the origin of komatiite-dominated greenstone volcanic rocks, TTG-type granitoids
195 and diapiric trondhjemites as well as the formation of the oldest cratonic core in the WDC
196 through assembly of micro-blocks and, in turn, bring constraints on the early earth
197 geodynamics and origin of Archean protocontinents.

198

199 **2. Regional geological and tectonic framework**

200 The Dharwar craton ([Fig.1](#)) preserves a large tilted section of Archean continental crust
201 displaying progressive transition from lower to upper continental crust from south to the north,
202 respectively, forming a wide time (3600-2500 Ma) window to early earth evolution (e.g., [Naqvi](#)
203 [and Rogers, 1987](#); [Jayananda et al., 2018](#)). This feature is unique to the WDC and makes it a
204 prime target to address crust formation processes, continental growth, mantle evolution, craton
205 building mechanisms and evolving tectonics of the Archean Earth. The craton comprises three
206 major lithological assemblages which include voluminous 3450-3200 Ma granitoids of
207 essentially TTG-affinity with much older (>3600 Ma) remnants, 2700-2600 Transitional TTGs,
208 two distinct groups of volcanic-sedimentary greenstone successions (>3200 Ma Sargur Group

209 and 3000-2600 Ma Dharwar Supergroup) and 3000-2520 Ma calc-alkaline to potassic plutons
210 (Jayananda et al 2018, 2019 and references therein). The Dharwar craton has been divided into
211 two sub-blocks (western and eastern Dharwar) based on the nature and abundance of
212 greenstones as well as the age of surrounding granitoids of TTG affinity, potassic granite
213 plutons, crustal thickness, and thermal records. The steep mylonitic shear zone along the
214 eastern margin of the Chitradurga greenstone belt (**Fig. 2**) has been considered as the
215 fundamental dividing line between these two cratonic blocks (Swami Nath et al., 1976) and a
216 paleosuture (Naqvi, 1985). Recent tectonic fabric data, metamorphic record, geochronological,
217 elemental and Nd isotope data show that the Dharwar craton contains three blocks (western,
218 central, and eastern) with independent thermal records and accretion histories (Peucat et al.,
219 2013; Jayananda et al., 2018). The western block has the oldest preserved TTG-greenstone
220 associations (3450-3200 Ma) with evidence for even older remnants (ca.3600 Ma; Meen et al.,
221 1992; Nutman et al., 1992; Bidyananda et al., 2011; Guitreau et al., 2017; Ravindran et al.,
222 2020) and two thermal events (3100 Ma and 2500 Ma; Jayananda et al., 2013a, 2015; Dasgupta
223 et al., 2019). In contrast, the central block has a mixture of old (3360-3000 Ma) and young
224 (2700-2520 Ma) which was involved in three thermal events (ca.3200 Ma, 2620 Ma and 2510
225 Ma; Jayananda et al., 2013a). Finally, the eastern block is mainly made of young crust of 2700
226 -2530 Ma that recorded only one thermal event close to 2510 Ma (Peucat et al., 2013). On the
227 cratonic scale, tectonic fabrics mapping, and kinematic analysis reveal dome and basin patterns
228 in the western block with widely spaced shear zones (Bouhallier et al.,1995; Chardon et al.,
229 2008), whereas the central and eastern blocks have spectacular flat fabrics linked to lateral
230 constrictional flow of hot orogenic crust with closely spaced shear zone network (Chardon et
231 al., 2011). Recent strain fabric data in combination with metamorphic facies, age zonation
232 patterns, Sm-Nd isotopes, and strong seismic reflectivity pointed towards Neoproterozoic craton
233 formation through assembly of the three crustal blocks through westward convergence of

234 oceanic lithosphere along the eastern margin of the Chitradurga greenstone belt ([Chadwick et](#)
235 [al., 2000](#); [Chardon et al., 2011](#); [Vijaya Rao et al., 2015](#); [Jayananda et al., 2013b, 2018, 2020](#)).

236 The Holenarsipur greenstone belt (HGB) and adjacent granitoids are considered to be
237 the oldest Archean crustal nucleus in the Dharwar craton ([Radhakrishna and Naqvi, 1986](#)).
238 Several studies on tectono-stratigraphy, strain patterns, metamorphic record, geochemistry and
239 the tectonic processes have significantly contributed to our fundamental understanding of
240 crustal architecture of the Hassan-Gorur-Holenarsipur region ([Chadwick et al., 1978](#);
241 [Ramakrishnan and Viswanatha, 1981](#); [Naqvi, 1981](#); [Hussain and Naqvi, 1983](#); [Naqvi et al.,](#)
242 [2009](#); [Bouhallier et al., 1993, 1995](#); [Chardon et al., 2008](#); [Jayananda et al., 2013a, 2015](#);
243 [Dasgupta et al., 2019](#)). Despite many studies, the stratigraphy of the Holenarsipur greenstone
244 belt and its temporal relationship with the surrounding granitoids (TTGs) is still debated.
245 [Ramakrishnan and Swami Nath \(1981\)](#) reported two distinct volcanic-sedimentary supracrustal
246 sequences comprising older Sargur Group (>3200 Ma) in the south and younger Bababudan
247 Group (<3000 Ma) to the north separated by the Tattekere conglomerate (e.g., [Chatterjee and](#)
248 [Das, 2004](#)). On the contrary, [Naqvi \(1981\)](#) argued that the Tattekere conglomerate does not
249 mark the separation between two stratigraphic sequences as cross-bedded quartzite between
250 two stratigraphic sequences itself does not form a proof for time break. Based on the map
251 patterns, [Naqvi \(1981\)](#) and [Hussain and Naqvi \(1983\)](#) have proposed that supracrustal rocks
252 predate the surrounding Peninsular gneisses while isotopic age data from [Monrad, \(1983\)](#) and
253 [Meen et al., \(1992\)](#) concluded that Peninsular gneisses are older. Zircon U-Pb dating from
254 felsic volcanic flow from higher stratigraphic level of ‘so called’ younger northern Bababudan
255 Group, yielded an age of 3298 ± 7 Ma ([Peucat et al., 1995](#)), indicating that the entire
256 Holenarsipur greenstone belt is ca.3300 Ma old, which in turn coeval with adjoining TTGs
257 ([Jayananda et al., 2015](#); [Guitreau et al., 2017](#)). According to [Drury et al. \(1987\)](#), the crustal
258 architecture of the Holenarsipur belt was shaped by large mushroom-type fold interferences

259 and earlier recumbent isoclinal folds refolded into upright folds. Subsequent strain fabrics
260 mapping and kinematic analysis by [Bouhallier et al. \(1993,1995\)](#) revealed classical crustal-
261 scale dome and keel patterns, typical of Archean cratons, wherein low density orthogneisses
262 define circular to elliptical domes whilst high-density greenstone sequences form keel
263 structures ([Gorman et al., 1978](#)). Between the interfering dome and keel structures, areas of
264 foliation triple points contain either ascending trondhjemite magma injections or down-going
265 high density ultramafic—mafic volcanic rocks which define vertical tectonites ([Bouhallier et
266 al., 1993](#)). However, recent phase equilibrium and mechanical modelling have argued that
267 sagduction was unlikely to have operated in the Archean ([Miocevich et al., 2022](#)). The HGB
268 was considered to have originated by migration of an oceanic crust from a spreading centre to
269 a trench forming an accretionary complex ([Kunugiza et al., 1996](#)).

270 Volcanic-sedimentary sequences stratigraphically equivalent to the Holenarsipur
271 greenstone basin are preserved in the adjacent regions like the Nuggihalli, Kalyadi, J.C. Pura,
272 Ghattihosahalli, Banasandra and Nagamangala belts ([see fig.2a](#)). These greenstone belts also
273 preserve voluminous komatiite-komatiitic basalt with minor basaltic rocks and interlayered
274 sedimentary rocks ([Jayananda et al., 2008, 2016; Maya et al., 2017; Tushipokla and Jayananda,
275 2013](#)). Published magmatic and detrital zircon ages together with Sm-Nd whole-rock isochrons
276 reveal major episodes of crust formation during 3600-3200 Ma ([Meen et al., 1992; Nutman et
277 al., 1992; Peucat et al., 1995; Hokada et al., 2013; Jayananda et al., 2008, 2015, 2019; Wang
278 and Santosh, 2019](#)). They form an extension of oldest cratonic nuclei preserved in the
279 Holenarsipur region. The cratonic core in the WDC experienced greenschist to amphibolite
280 facies metamorphism that affected the whole Dharwar craton close to 2500 Ma with an earlier
281 thermal event ca. 3100 Ma ([Bouhallier, 1995; Jayananda et al., 2013a 2015; Dasgupta et al.,
282 2019](#)).

283

284 3. Lithological assemblages, field relationships and petrography

285 The cratonic core in the WDC (**Fig.2a**) comprises the komatiite dominated Sargur Group
286 volcanic-sedimentary greenstone sequences and adjoining granitoids (TTGs) which, in turn,
287 were intruded by ca. 3200 Ma diapiric trondhjemites. It shows classical dome and keel patterns
288 typical of oldest cratonic cores in the Archean cratons ([Bouhallier et al., 1993, 1995](#);
289 [Choukroune et al., 1995](#); [Chardon et al., 1996, 1998, 2008](#)). The Holenarsipur greenstone belt
290 together with other stratigraphically equivalent volcanic-sedimentary sequences found in the
291 J.C.Pura, Nuggihalli, Kalyadi, Banasandra and Nagamangala greenstone belts and adjoining
292 TTGs form the cratonic core in the Western Dharwar Craton ([Chardon, 1997](#); [Ramakrishnan](#)
293 [et al., 1994](#); [Nutman et al., 1992](#); [Meen et al., 1992](#); [Peucat et al., 1995](#); [Guitreau et al., 2017](#);
294 [Jayananda et al., 2008, 2015](#); [Hokada et al., 2013](#); [Bidyananda et al., 2016](#); [Lancaster et al.,](#)
295 [2015](#); [Wang and Santosh, 2019](#)). Here we present our new field data on the Holenarsipur
296 greenstone belt and adjacent granitoids together with published data of the other
297 stratigraphically equivalent greenstone belts ([Swami Nath and Ramakrishnan, 1981](#);
298 [Radhakrishna, 1983](#)) in the cratonic core and their adjacent granitoids of TTG-affinity.

299 The Holenarsipur greenstone belt (**Fig. 2b**) and surrounding granitoids (TTGs) display
300 complex internal architecture ([Naqvi, 1981](#); [Ramakrishnan and Viswanatha, 1981](#); [Bouhallier](#)
301 [et al., 1993, 1995](#); [Choukroune et al., 1995, 1997](#)). The volcano-sedimentary greenstone
302 sequence comprises abundant ultramafic assemblages (komatiite to komatiitic basalts),
303 amphibolite with interlayered northwest – southeast trending quartzite-conglomerate and
304 pelites in the central part separating the greenstone into three units. Banded Iron Formations
305 (BIFs) are confined to the summits of the north - south trending eastern most part of the
306 greenstone belt (**Fig.2c**). Diapiric trondhjemites (3230-3170 Ma; [Jayananda et al., 2015](#);
307 [Guitreau et al., 2017](#)) intruded the greenstone sequences and adjoining granitoids whilst 3150

308 Ma granite veins intrude the TTG gneisses in the Southwestern part. Minor anorthosite (3240
309 Ma; Panicker et al., 2021) intrude the greenstone sequences in the south-eastern part.

310 Our extensive field work revealed the composite nature of the Holenarsipur greenstone
311 belt which comprises three distinct tectonic units identified based on distinct lithological
312 assemblages and their corresponding tectonic features. We labelled these units Southwestern,
313 Northcentral, and Eastern blocks (see fig.2b, c) and describe them below.

314

315 3.1 Southwestern block

316 The southwestern block (see fig.2b) comprises dominant metamorphosed ultramafic
317 (serpentinite ±olivine bearing dunite, serpentine-talc-tremolite schist, talc-chlorite schist,
318 actinolite-tremolite schist) with minor mafic volcanic rocks (plagioclase-chlorite-actinolite-
319 hornblende schist) associated with sedimentary rocks (conglomerate, quartzite, garnet-mica-
320 staurolite-kyanite bearing pelites) and surrounding gneisses (granitoids of TTG affinity).

321 The ultramafic-mafic rocks are found as either large discontinuous bands or
322 dismembered outcrops distributed within the TTG-gneisses (Fig. 3.1a). Their contacts with
323 adjacent granitoids (TTG-type gneisses) and sedimentary rock sequences are tectonized. In the
324 southernmost part, the ultramafic rocks are found as numerous dismembered fragments or
325 outcrops. A gradual increase in their abundance can be followed from disrupted fragments in
326 the southernmost part to continuous outcrops to the northern part of the southwestern block.
327 These ultramafic rocks were disrupted and fragmented probably during the intrusion of
328 trondhjemite magmas. They occasionally show crude pillow structure with chilled margins
329 which indicate their eruption in a marine environment (Fig. 3.1b). The ultramafic rocks are
330 medium to fine grained containing serpentine-talc-tremolite or tremolite-actinolite, and their
331 primary mineralogy is nearly absent except for remnants of olivine and/or clinopyroxene rarely
332 preserved in a few samples (Fig. 4.1a). The mafic rocks are found as large discontinuous

333 outcrops or continuous bands which do not show contact with ultramafic units. They are
334 medium- to fine-grained with abundant actinolite, tremolite, minor plagioclase with rare
335 hornblende. This mineralogy developed during post-magmatic hydrothermal alteration or
336 metamorphism. These ultramafic-mafic assemblages together with granitoids (TTG- gneisses)
337 are bounded by southeast to northwest trending shallow water sedimentary successions
338 (conglomerate-quartzite-pelites) which define the northern limit of the southwestern block ([see](#)
339 [fig. 2c](#)). Among the sedimentary sequences, conglomerates contain rounded to stretched
340 ellipsoidal quartz pebbles embedded in quartz-rich matrix ([Fig. 3.1c](#)) and quartzites are often
341 fuchsite-bearing and display ripple marks ([Fig. 3.1d](#)) and crossbedding ([Fig.3.1e](#))
342 characteristic of shallow shelf environment. Pelites comprise diverse assemblages including
343 garnet-biotite, garnet-biotite-muscovite-kyanite, garnet-biotite-staurolite-kyanite bearing
344 assemblages ([Fig.3.1f](#)) indicating that equilibration took place at 4-7 kbars and ~550°C
345 ([Bouhallier, 1994; Dasgupta et al., 2019](#)). Detrital zircons from garnet-mica-kyanite-bearing
346 pelites yielded ages ranging from 3560 to 3230 Ma ([Nutman et al., 1992](#)). The contact between
347 the ultramafic-mafic rocks and adjoining granitoids are tectonized. The dominant ultramafic
348 with minor mafic rock assemblages with crude pillows point to a possible oceanic plateau.

349 The adjacent granitoids of TTG affinity comprise several associations of grey to dark
350 grey granodioritic to tonalitic gneisses containing the remnants of whitish grey granite to
351 trondhjemite, intermediate to mafic layers. The older remnants of whitish grey granite to
352 trondhjemite are found as fragments within tonalite or granodiorite layers ([Fig.3.1g](#)). They
353 contain quartz, plagioclase (An₁₀₋₁₂), minor microcline, biotite and occasional primary
354 muscovite. The dark grey tonalite layers are abundant and often show diffusive contacts with
355 granodiorite suggesting their coeval nature. Tonalite contains quartz, plagioclase (An₁₅₋₂₆),
356 minor microcline, hornblende, biotite with accessory zircon, apatite, titanite and opaques. The
357 voluminous grey granodiorite contains older remnants of biotite-rich mafic enclaves

358 (Fig.3.1h). They have quartz, plagioclase (An₁₆₋₂₂), microcline, biotite in their mineral
359 assemblage. Both tonalite and granodiorite were intruded by veins of trondhjemite which are
360 often folded (Fig.3.1i). These trondhjemite veins contain quartz, sodic plagioclase (An₁₀₋₁₃),
361 minor microcline, biotite and rare muscovite. Late (3150 Ma; Jayananda et al., 2015)
362 granodiorite to granite veins and/or dykes intrudes TTGs. Trondhjemite injects into the
363 gneisses and form vertical tectonites at foliation triple points (Fig.3.1j).

364

365 3.2 Northcentral block

366 The greenstone sequence of the Northcentral block is surrounded by TTG-type
367 granitoids and trondhjemite intrusions on its northern and western margins. This greenstone
368 unit was probably dismantled into two fragments (see fig.2b) during trondhjemite (Halekote-
369 type) intrusion. The greenstone assemblages comprise ultramafic to mafic volcanic rocks with
370 minor felsic pyroclastic flows and interlayered sedimentary rocks (garnet-chlorite-chloritoid-
371 bearing pelite). Ultramafic-mafic rocks erupted in marine environment as revealed by flow top
372 pillow breccias (Fig. 3.2a). They show constrictional fabrics forming vertical tectonites at
373 foliation triple junctions developed at the junction of interfering rising gneiss domes and down
374 going ultramafic volcanics (Bouhallier et al., 1993). The ultramafic volcanic rocks contain
375 serpentine-talc-tremolite or talc-tremolite-actinolite or talc-chlorite whilst mafic volcanic rocks
376 exhibit actinolite-tremolite with minor plagioclase or actinolite-hornblende-plagioclase in the
377 mineral assemblage (Fig. 4.1b, c). Felsic volcanic rocks (quartz- sodic plagioclase - minor k-
378 feldspar-biotite) are confined to the southwestern margin of Northcentral block. This felsic
379 volcanic unit corresponds to higher stratigraphic levels that comprises pyroclastic flows
380 erupted in sub-aerial conditions as revealed by blue quartz phenocrysts, medium grained matrix
381 and tiny fine-grained angular rock fragments within fine grained ash (Fig. 3.2b). Chlorite-
382 chloritoid-garnet-bearing pelitic sedimentary rocks (Fig. 3.2c) are found along the southern-

383 margin close to the contact with the Southwestern block. The mineralogy of these chloritoid-
384 bearing pelitic rocks imply ultramafic to mafic provenance and deposition in an oceanic
385 environment far from a sizable continent. The dominant ultramafic-mafic assemblages with
386 minor felsic volcanic rocks and associated chloritoid-chlorite-bearing pelite rocks point to an
387 ocean island arc spatially associated with a volcanic plateau.

388 The greenstone volcanic rocks show the imprints of contact metamorphism along their
389 boundary with ca. 3200 Ma diapiric Halekote-type trondhjemite as revealed by large randomly
390 oriented crystals of actinolite, tremolite and hornblende in mafic-ultramafic volcanic rocks
391 (**Fig.3.2d**). Numerous quartz-veins with large tourmaline crystals also injected into the
392 greenstone volcanic sequences along the contact zone with diapiric trondhjemite.

393 The granitoids around the Northcentral block is dominantly coarse- to medium- grained
394 dark grey tonalitic to grey granodioritic gneisses defining the domal pattern (**Bouhallier et al.,**
395 **1993**). These gneisses are traversed by medium-grained whitish grey trondhjemite veins and
396 highly coarse-grained pegmatites. The contacts between greenstone and granitoids (TTG-type)
397 are tectonized. The tonalitic gneisses contain quartz, plagioclase (An_{17-25}), biotite, minor
398 microcline with biotite and hornblende whilst granodiorite has quartz, plagioclase (An_{15-20}),
399 microcline and biotite with common accessory phases such as zircon, apatite, titanite and
400 opaques.

401

402 3.3 Eastern block

403 The eastern block forms a thin (500 m to 3 km width) elongated band that runs for about
404 60 km in north-south direction (**see fig. 2b**). The eastern block and adjacent granitoids (TTG-
405 type) experienced strong deformation associated with regional shortening that affected the
406 whole Archean crust in the Dharwar craton close to 2500 Ma (**Bouhallier et al., 1993**). In the
407 northern part of the Holenarsipur greenstone belt, an E-W road cut displays a complete section

408 of volcano-sedimentary sequence characteristic of the Eastern block (**Fig.3.3a**). From east to
409 west, the greenstone sequence begins with pillowed mafic-ultramafic rocks with chilled
410 margins followed by phyllite, chert layers, BIFs, several fine-grained sheeted mafic dykes,
411 plagiogranite, gabbros to layered gabbros, fine-grained ultramafic rocks to medium-grained
412 peridotite to the westernmost part (**Fig.3.3b**). Further west, peridotitic rocks show tectonic
413 contacts with the adjacent granitoids of TTG affinity. The pillowed ultramafic rocks contain
414 serpentine-talc-tremolite, talc-tremolite-actinolite or talc-chlorite whilst mafic volcanic rocks
415 display chlorite-actinolite-plagioclase in the mineral assemblages. The gabbros and layered
416 gabbros show plagioclase-hornblende with remnants of clinopyroxene. The plagiogranite
417 exhibit dominant sodic plagioclase with minor quartz and rare biotite. The medium-grained
418 peridotitic ultramafic rocks in the western margin contain serpentine with remnants of olivine,
419 serpentine-talc-tremolite in their mineralogical assemblage. Immediately south of the road cut,
420 a thin band of intraformational conglomerate interlayered with phyllite and BIFs was
421 documented by [Swami Nath and Ramakrishnan \(1981\)](#). The conglomerate contains flat to
422 angular pebbles of chert, fuchsite-bearing quartzite, and mafic volcanic rocks with greywacke
423 matrix suggest fragmentation within the basin without much transportation. The southern part
424 of the eastern block corresponds to deeper levels that exhibit serpentine-talc-tremolite and
425 actinolite-hornblende schist (**Fig. 4.1d**) as well as gabbro-anorthosite, plagiogranite, and
426 peridotitic rocks along an east-west section. These field evidence include rock sequences like
427 peridotite-foliated gabbro - plagiogranite - sheeted dykes-pillowed ultramafic volcanics with
428 chert and BIFs along E-W cross-sections at different crustal levels which suggest that the
429 eastern block could probably correspond to a titled and deformed panel of preserved oceanic
430 crust close to an oceanic spreading centre (**Fig. 3.3b, c**).

431 The granitoids (TTG-gneisses) adjoining to the eastern block comprise dark grey
432 tonalitic to grey granodioritic gneisses, which contain thin whitish trondhjemite veins along

433 steep foliation. These TTG-type granitoids are traversed by strong transcurrent ductile to
434 brittle-ductile shears which locally reoriented foliation and lineation (Bouhallier et al., 1993).
435 The tonalite has quartz, plagioclase (An₁₉₋₂₇), traces of microcline, hornblende, biotite whilst
436 granodiorite contain quartz, plagioclase (An₁₄₋₂₀), subordinate microcline, biotite, occasional
437 hornblende, and common accessory minerals such as zircon, apatite, titanite and opaque
438 phases. Whitish grey trondhjemite veins are found along the steep foliation of tonalitic to
439 granodioritic gneisses contain minor quartz and abundant sodic plagioclase (An₁₀₋₁₂), minor
440 microcline with rare biotite and muscovite.

441

442 3.4 Diapiric trondhjemite intrusions

443 Diapiric trondhjemites intruded as plutons at the contact zone between the three
444 greenstone units and the TTG-type granitoids (see fig.2b, c) but they also occur as veins and
445 sheets (often folded) within the adjoining TTGs away from main intrusions. They also define
446 foliation triple points at junctions of interfering dome-keel structures forming vertical tectonites
447 (see fig.3.1j). These diapiric trondhjemite intrusions are dominantly homogeneous coarse-
448 grained trondhjemite with minor granite in their core. They have quartz, plagioclase (An₁₀₋₁₂),
449 minor microcline, biotite with or without muscovite, accessory zircon, titanite and Fe-Ti
450 oxides. Detailed tectonic fabric analysis reveals that the intrusion of diapiric trondhjemites
451 shaped the fundamental architecture of the preserved crustal panel in the Holenarsipur
452 greenstone belt (Bouhallier et al., 1993).

453

454 3.5 Lithological assemblages along the contact zone of the three tectonic units

455 The contact zone between the three tectonic units is key to understand the crustal
456 architecture, tectonic relationships between the three greenstone units. Distinct sedimentary
457 assemblages are found along the northern and southern margin of the contact zone between

458 southwestern and northcentral block (see fig. 2b-c). Along the northern limit of the
459 southwestern block, thick quartzite with crossbedding (see fig. 3.1e) and ripple marks (see fig.
460 3.1d), following quartz-pebble conglomerate (see fig. 3.1c), mark a shallow shelf environment.
461 These quartzite-conglomerate layers are followed by quartz-mica-kyanite-staurolite-garnet-
462 bearing pelite (see fig. 3.1f) which spatially associated with ultramafic rocks and granitoids
463 (TTGs) suggesting a mixed provenance. On the contrary mineralogy (chlorite-chloritoid-
464 garnet) and major element composition of the pelite unit interlayered with greenstone volcanic
465 rocks (Bouhallier, 1995) along the south-eastern margin of the northcentral block (see fig. 2c)
466 suggests its derivation from dominant ultramafic-mafic source. Within the eastern block, close
467 to the contact with the northcentral block (see fig.2b), a polymict conglomerate with sparse
468 angular fragments of greenstone volcanic rocks/sedimentary rocks in clay matrix
469 (Ramakrishnan and Viswanatha, 1981) indicate local fragmentation within the oceanic basin.
470 The contact zone of the three blocks is marked by the intrusion of diapiric trondhjemites
471 wherein numerous quartz veins that contain large crystals of tourmaline injected into volcanic-
472 sedimentary sequences. The origin of tourmaline-bearing quartz veins is probably linked to the
473 volatile-rich fluid flow associated with the intrusion of ca. 3200 Ma diapiric trondhjemite as
474 also suggested by Rogers et al. (1986). To summarize, the observed field relationships along
475 the contact zone imply that these three greenstone units representing three tectonic units
476 probably originated in distinct settings and their juxtaposition coincided with the formation of
477 diapiric trondhjemites.

478

479

480 3.6 Stratigraphically equivalent greenstone belts and adjacent granitoids rocks in the 481 cratonic core

482

483 The stratigraphically equivalent greenstone sequences to the Holenarsipur greenstone
484 belt include Nuggihalli, Kalyadi, J.C. Pura, Ghattihosahalli, Banasandra and Nagamangala
485 greenstone belts (see fig. 2a). These greenstone belts are surrounded by ca.3350-3300 Ma
486 granitoids rocks which were intruded by ca. 3200-3150 Ma diapiric trondhjemites, as well as
487 3000 Ma and 2600 Ma potassic granites (Meen et al., 1992; Chardon, 1997; Jayananda et al.,
488 2006, 2019; Ao et al., 2021).

489
490 *3.6.1 The Nuggihalli greenstone belt*

491 The Nuggihalli greenstone belt is an elongated NNW-SSE trending volcano-sedimentary
492 sequence that extend about 50 km from Kempinakote in the south to Arsikere in the north (Jafri
493 et al., 1983). The greenstone belt is surrounded by granitoids which comprise TTG-type
494 banded to migmatitic gneisses, and their contacts have been tectonized. Nuggihalli greenstone
495 stratigraphically correlates with the Holenarsipur greenstone belt (Swami Nath and
496 Ramakrishnan, 1981). It is difficult to establish relationships among the lithologies as
497 lithological assemblages were disrupted during strong shear deformation. The dominant
498 lithological sequences include metamorphosed ultramafic rocks (peridotite, serpentinite and
499 dunite with chromite layers, talc-tremolite schists; Fig. 5.1a), mafic rocks (gabbros and garnet-
500 amphibolite), anorthosites, granite veins (plagiogranite?) and titaniferous magnetite.
501 Metasedimentary rocks such as fuchsite-bearing quartzite, garnet-mica (\pm kyanite or staurolite)
502 metapelite were documented in central and northern part within granitoids (Swami Nath and
503 Ramakrishnan, 1981; Jafri et al., 1983). The ultramafic rocks can be found as lenses and
504 elongated bodies throughout the belt. Peridotitic komatiites with chromite layers are well
505 preserved at Jambur, Nuggihalli, Tagadur, Byrapur and Bhaktarahalli chromite mines. On a
506 regional scale, the greenstone belt defines NNW trending fabrics and was affected by strong
507 shear deformation. Whole-rock Sm-Nd isotope data of Nuggihalli volcanic rocks together with

508 data from the adjoining Kalyadi belt define an imprecise regression age of 3284 ± 310 Ma
509 (Jayananda et al., 2008). The Nuggihalli greenstone belt was equilibrated in greenschist to
510 amphibolite facies. An increase of metamorphic grade can actually be followed from south to
511 north as documented by Jafri et al. (1983). The lithological association of dominant ultramafic-
512 mafic volcanic rocks with chromite layers, gabbro-anorthosites and granite (plagiogranite?)
513 injections point to a preserved section of oceanic crust close to spreading centre. The komatiites
514 contain serpentine, tremolite, talc, actinolite, chlorite (Fig. 5.2a), chromite with rare olivine
515 and enstatite (Jayananda et al., 2008).

516

517 3.6.2 Kalyadi greenstone belt

518 The volcanic-sedimentary sequences in the Kalyadi region crop out as several large
519 discontinuous boudins or small 1-2 km bands within the TTG-type granitoids which in turn
520 intruded by diapiric trondhjemites. The contacts between the greenstone sequences and
521 surrounding TTGs are tectonized. Late potassic granites (ca. 2620 Ma Arsikere-Banavara
522 plutons) are intrusive into the Kalyadi greenstone belt and adjoining TTGs (Meen et al., 1992;
523 Jayananda et al., 2006). The greenstone sequences are dominated by komatiitic volcanic rocks
524 which include serpentinite, serpentine-talc-tremolite and talc-chlorite schists (Fig. 5.2b).
525 Minor amphibolites and hornblende schists are associated with ultramafic rocks (Fig. 5.1b).
526 Nodular structures found in komatiite comprise spherical nodules with 1-2 cm diameter
527 surrounded by fine grained matrix (Jayananda et al., 2008). The minor sedimentary
528 assemblages include cherty quartzites and metapelites interlayered with volcanic rocks (Subba
529 Rao and Naqvi, 1999) and immediately north fuchsite-bearing quartzite that contain chromitite
530 layers (Swami Nath and Ramakrishnan, 1981). The abundant komatiites and associated minor
531 sedimentary rocks point to their eruption in a marine environment representing a possible
532 section of volcanic plateau fragment spatially associated with an oceanic spreading centre.

533 Published U-Pb zircon age for the granitoids adjacent to the Kalyadi greenstone belt is
534 rather scanty. However, [Meen et al \(1992\)](#) have estimated an age of 3390 Ma for the magmatic
535 protoliths of low-Rb gneisses south of the Kalyadi belt. U-Pb zircon ages of tonalitic to
536 granodioritic gneisses northeast of the Kalyadi belt indicate 3310-3205 Ma ([Chardon, 1997](#))
537 whilst detrital zircons from chromite-bearing fuchsite indicate 3230 Ma ([Nutman et al., 1992](#)),
538 and directly ~30 km to the west, granitoids of TTG-affinity in the southern boundary of
539 Bababudan basin provide an age range of 3350-3270 Ma ([Jayananda et al., 2015](#)).

540

541 *3.6.3 J.C. Pura greenstone belt*

542 The J.C. Pura greenstone belt preserves voluminous fresh ultramafic volcanic sequences of
543 komatiite and komatiitic basalt composition which are interlayered with minor cherty quartzites
544 ([Venkata Dasu et al., 1991](#); [Ramakrishnan et al., 1994](#); [Jayananda et al 2008, 2016](#)). Komatiites
545 form spectacular pillows, pillow breccias and flowtop breccia with chilled margins ([Jayananda](#)
546 [et al., 2016](#)). Crude spinifex structures were also documented ([Venkata Dasu et al., 1991](#)).
547 Thick flows (70-90 m) of undifferentiated komatiites form prominent ridges around Rampura-
548 Sasivala region. These rocks originated as long drawn lava lobes and lava sheets spread over
549 tens of metres ([Jayananda et al., 2016](#)). The lithological association of voluminous komatiite,
550 komatiitic basalts and minor basalt with flowtop pillow breccias (**Fig. 5.1c & d**) imply their
551 eruption in marine environment possibly as an oceanic plateau. Komatiites define an Sm-Nd
552 whole-rock regression age of 3384 ± 200 Ma ([Jayananda et al., 2008](#)) whilst detrital zircons
553 from associated quartzites yield an age 3230 Ma ([Ramakrishnan et al., 1994](#)). Komatiites
554 contain serpentine, serpentine-talc-tremolite, talc-chlorite with accessory chromite and spinel.
555 Remnants of olivine are found in serpentine-rich layers (**Fig. 5.2c**). The adjacent granitoids
556 defines circular to sub-circular domes ([Chardon et al., 1996](#)) and provide U-Pb zircon ages of
557 3315-3217 Ma ([Chardon, 1997](#); [Jayananda unpublished data](#)) whilst trondhjemite to

558 granodioritic intrusions into the TTG-type granitoids indicate U-Pb zircon ages close to 3198
559 Ma (Chardon, 1997).

560

561 3.6.4 *Ghattihosahalli greenstone belt*

562 The Ghattihosahalli belt forms a small and elongated band along the western boundary of the
563 ca. 2700 Ma Chitradurga greenstone belt. Komatiites are the most abundant lithologies which
564 are associated with minor amphibolites, fuchsite-bearing quartzites which are interbedded with
565 barite layers (Radhakrishna and Srinivasaiah, 1974). These komatiites occur as bouldery
566 outcrops on the hill to the south of Kummanghatta and occasionally show relict spinifex texture
567 (Viswanatha et al 1977). Petrographic study reveals abundant serpentine (>90%) in the
568 majority of samples and serpentine-talc-tremolite in a few samples (Fig. 5.2d). The dominant
569 komatiitic volcanic rocks together with fuchsite-bearing quartzite and barite layers point to
570 remnants of an oceanic plateau.

571 The granitoids adjacent to the Ghattihosahalli belt comprises banded dark grey tonalitic
572 to grey granodiorite with several leucocratic aplitic veins along the foliation. SIMS U-Pb zircon
573 ages provide ages of 3318 ± 15 Ma to 3259 ± 7 Ma for the TTG-type granitoids which in turn
574 intruded by 3000 Ma potassic Hosdurga granite (Jayananda et al. in prep).

575 3.6.5 *Banasandra greenstone belt*

576 The Banasandra greenstone belt is located on the southwestern margin of the Chitradurga
577 greenstone belt (see fig. 2a). This belt occurs as a NW-SE trending bands which comprises
578 well-preserved voluminous peridotitic komatiites around Banasandra, Kodihalli and
579 Kunikenahalli (Srikantia and Bose, 1985). These komatiites are fine-grained forming pillow
580 breccias or flowtop breccias (Jayananda et al., 2016) in the south near Kunikenahalli and
581 display spectacular spinifex structures (Fig.5.1e) to the north close to Kodihalli village
582 (Jayananda et al., 2008, 2016). Pillow breccias and flowtop breccias are considered to underlie

583 the spinifex (Srikantia and Bose, 1985). The studied komatiites consist of serpentine
584 (cumulate), serpentine-talc-tremolite, talc-tremolite and talc-chlorite in mineral assemblages
585 (Fig. 5.2e). Serpentine-tremolite-talc and talc-chlorite assemblages indicate low grade
586 metamorphism under greenschist facies conditions. The discontinuous outcrop patterns
587 together with dominant komatiite volcanic rocks, which form spectacular pillow breccia and a
588 spinifex texture, imply that Banasandra komatiites correspond to possible remnants of an
589 oceanic plateau.

590 The adjacent granitoids are dominantly dark grey banded tonalitic to granodioritic
591 gneisses which are intruded by whitish grey trondhjemite. Zircons from tonalitic to
592 granodioritic layers indicate U-Pb ages of 3310 ± 15 Ma and 3210 ± 10 Ma (Jayananda et al.,
593 2019). All of which, in turn intruded by ca. 3000 Ma high-K granite plutons (Jayananda et al.,
594 2019; Ao et al., 2021).

595

596 3.6.6 Nagamangala greenstone belt

597 The Nagamangala greenstone belt is located in the easternmost part of the WDC (see
598 fig. 2a) and it extends about 30 km from Nelligere in the north to Honnakere in the south
599 (Srikantia and Venkataramana, 1989). The greenstone belt is surrounded by ca. 3300-3200 Ma
600 granitoids, all of which were intruded by ca. 3000 Ma potassic plutons (Jayananda et al., 2019).
601 The preserved volcanic rocks include abundant ultramafic rocks of komatiite affinity with
602 minor basalts which are interlayered with sedimentary rocks (Fig. 5.1f) (pelite-quartzite-
603 carbonate). The ultramafic rocks display spinifex and pillow structures whilst mafic rocks
604 forming pillows with chilled margins (Devapriyan et al., 1994; Tushipokla and Jayananda,
605 2013). The ultramafic volcanic rocks contain serpentine-tremolite-talc and tremolite-actinolite
606 whilst mafic volcanic rocks exhibit actinolite-hornblende-plagioclase assemblage (Fig. 5.2f).
607 The lithological associations of dominant ultramafic to minor mafic volcanic rocks show pillow

608 structures with interlayered quartzite and pelite which possibility imply an eruption in marine
609 environments alike oceanic plateaus. Mineral assemblages indicate greenschist to lower
610 amphibolite layers metamorphic conditions (Tushipokla and Jayananda, 2013).

611 The surrounding granitoids comprise strongly banded tonalitic to granodiorite which
612 contain trondhjemite veins along the foliation which are often boudinaged (Fig. 6.1a-f). The
613 gneisses are affected by strong deformation associated with regional shortening and steep
614 foliation. Zircons from tonalitic to granodioritic grey gneisses yield U-Pb zircon ages of 3205
615 ± 5 to 3310 ± 10 Ma (Jayananda et al., 2019) whilst high-potassic plutons emplaced along TTG
616 - greenstone contact zone provide U-Pb zircon ages of 2989-2974 Ma and 2620-2600 Ma
617 (Jayananda et al. 2006, 2019).

618 To summarize stratigraphically equivalent Sargur Group (ca.3400-3200 Ma)
619 greenstone volcanic sequences in the cratonic core are dominated by komatiite-komatiitic
620 basalts with minor basaltic rocks. The primary mineralogy of volcanic rocks is absent (except
621 rare olivine and orthopyroxene) and observed mineralogy developed during hydrothermal
622 alteration and low-grade metamorphism. The volcanic rocks are interlayered with shallow
623 water shelf sedimentary sequences like quartzite-pelite-carbonate-BIFs. The contacts between
624 greenstone sequences and adjoining granitoid rocks are tectonized. Our field observations
625 combined with published tectonic fabrics analyses in the WSW-ENE crustal corridor in the
626 cratonic core (Bouhallier et al., 1993, 1995; Chardon et al., 1996, 1998, 2008) reveal dome and
627 keel architecture with major shear zones traverse the eastern block of the Holenarsipur,
628 Nuggihalli-Kalyadi, and Nagamangala greenstone belts (Fig. 6.1g).

629

630 **3.7 Structural patterns**

631 Dome and basin structures are a fundamental feature of TTG-greenstone terranes of many
632 oldest Archean cratons (Gorman et al., 1978; Bouhallier et al., 1993; Chardon et al., 1996,

633 2009; Choukroune et al., 1995, 1997; Rey et al., 2003; Whitney et al 2004). These dome and
634 basin patterns were observed in different spatial scales and their origin have been attributed to
635 various processes operated in diverse tectonic settings (Bouhallier et al., 1995; Collins et al.,
636 1998; Hamilton, 1998; Kusky et al., 2021; Zulauf et al., 2019). Structural studies and seismic
637 tomographic data on the western Dharwar craton reveal a first order cratonic internal
638 architecture that comprises thick crust (40-45 km) with dome and basin patterns and crustal
639 scale shear zone network (Drury and Holt, 1980; Chadwick et al., 1978, 1981; Gupta et al.,
640 2003; Bouhallier et al., 1995; Chardon et al., 1998, 2008). Detailed structural analysis based
641 on combined satellite-based remote sensing data (Spot and LANDSAT image analysis)
642 together field-based strain fabrics measurements in the cratonic core reveal crustal scale
643 dome and basin patterns bounded by widely spaced shear zone network (Bouhallier et al.,
644 1995; Chardon et al., 1996, 2008; Choukroune et al., 1997). In the NE part of the cratonic
645 core, a detailed structural analysis involving strain fabrics measurements and kinematic
646 analysis around the ca.3400 Ma J.C. Pura greenstone belt and ca.3300 Ma adjacent granitoids
647 indicate the development of diapiric structures due to vertical displacements of high-density
648 greenstone volcanics associated with gravitational instabilities (Chardon et al., 1996). This
649 diapiric event is possibly linked to the emplacement of trondhjemite to granodiorite plutons
650 close to 3200-3150 Ma (Meen et al.,1992; Chardon, 1997; Jayananda et al., in prep.).
651 Foliation trajectories together with strain variation and kinematic analysis of the Holenarsipur
652 greenstone belt and adjoining granitoid gneisses reveal that two deformation events have
653 shaped the crustal architecture of the region. Kinematic indicators along the contact between
654 greenstone and gneisses indicate downward displacement of high-density greenstone
655 assemblages relative to granitoids (Bouhallier et al., 1993). The regional diapiric event is
656 contemporaneous with the intrusion of ca.3200 Ma Halekote trondhjemite which produced
657 circular to elliptical domes surrounded by greenstone sequences. The foliation triple points

658 marked by vertical tectonics developed at junction of converging lineation associated with
659 interfering domes and keel pattern (Bouhallier et al., 1993; Choukroune et al., 1997).

660 Sinistral ductile shearing affected the eastern block (see fig.2b) of the Holenarsipur belt and
661 adjacent gneisses which is interpreted as a result of vertical movements linked to body forces
662 with minor horizontal shortening (Bouhallier et al. 1993). Further to the south in Gundlupet
663 region, Bouhallier et al (1995) presented a 3D geometrical information of diapiric structures
664 from a tilted crust in the WDC corresponding to various depths developed elliptical gneiss
665 domes and greenstone keel with foliation triple points at the junction of converging lineation
666 (Bouhallier et al., 1995; Choukroune et al., 1997).

667 In summary our field observations together with published strain fabrics data including
668 foliation trajectories and kinematic indicators in the WSW – ENE corridor of the cratonic core
669 (Fig. 6.1g) reveal circular to elliptical dome-keel patterns (Bouhallier et al., 1993,1995;
670 Choukroune et al., 1997; Chardon et al., 1996, 2008). Kinematic indicators reveal downward
671 displacement of high-density greenstone with respect to adjacent low-density TTG-type
672 granitoids. Foliation triple points defined by vertical tectonites are associated with regions of
673 subsidence of high-density greenstone assemblages. These dome and basin structures are also
674 compatible with experimental analogue modelling (Bouhallier et al., 1995). Late transcurrent
675 shear zones are confined to the eastern boundary Holenarsipur and further south to the
676 Gundlupet region and Banasandra and Nagamangala greenstone belts (see fig. 2) which partly
677 reorient the foliation and lineation of dome-keel patterns (Bouhallier et al., 1995; Chardon et
678 al., 2008). The dome and keel patterns of the cratonic core are compatible with the dome and
679 basin structures documented in the cratonic cores in the East Pilbara and Kaapvaal cratons
680 (Anhaeusser, 2014; Collins et al., 1998; van Kranendonk et al., 2015).

681

682 4. Metamorphic record

683 Metamorphic pressure-temperature history of the cratonic core in WDC is poorly known except
684 for a few studies by [Bouhallier \(1995\)](#), [Jayananda et al \(2013a\)](#) and [Dasgupta et al \(2019\)](#) on
685 the Holenarsipur region. Studies on the regional metamorphism of the WDC reveal a
686 progressive increase in metamorphic grade from north to south ([Raase et al., 1986](#); [Jayananda
687 et al., 2013a, 2020](#) and references therein). Mineral assemblages (chlorite-actinolite-tremolite-
688 serpentine or chlorite-actinolite-garnet) in the remnants of the Sargur Group greenstones from
689 the southern margin of the Bababudan basin indicate greenschist facies ([Chadwick et al., 1981](#))
690 whilst garnet-staurolite-kyanite-chloritoid-bearing pelites in the Holenarsipur region indicate
691 amphibolite facies conditions ([Bouhallier et al., 1995](#); [Dasgupta et al., 2019](#)). Further south in
692 the Sargur-Gundlupet region hornblende-clinopyroxene bearing mafic rocks and appearance
693 of orthopyroxene in migmatitic gneisses marks amphibolite to granulite facies transition
694 ([Bouhallier et al., 1995](#); [Jayananda et al., 2013a](#)). The documented progressive increase of
695 metamorphic grade from north to south is related to the end Archean (ca.2500 Ma)
696 tectonothermal event (“hot orogen”) that affected the whole Archean crust of the Dharwar
697 craton ([Peucat et al., 2013](#); [Chardon et al., 2011](#); [Jayananda et al., 2013a, 2018](#)). Evidence for
698 an earlier episode of metamorphism close to ca. 3100 Ma is preserved in the Holenarsipur
699 region ([Jayananda et al., 2013a, 2015](#); [Dasgupta et al., 2019](#)). To summarize, the cratonic core
700 in WDC was affected by two major thermal events during ca. 3100 Ma and 2500 Ma
701 ([Jayananda et al., 2013a](#); [Dasgupta et al., 2019](#)). The 3100 Ma event was immediately preceded
702 by a major tectono-magmatic event marked by the intrusion of hot trondhjemite magmas,
703 partial convective overturn of mid to lower crust, and formation of dome-keel pattern.

704

705 **5. Analytical methods**

706 *5.1 Elemental geochemistry*

707 We present new major and trace elements data for the Holenarsipur greenstone volcanic
708 rocks ([Supplementary Table 1](#)), spatially associated granitoids (TTG-type gneisses) and
709 diapiric trondhjemites ([Supplementary Table 2](#)), as well as new LA-ICP-MS U-Pb zircon ages
710 ([Supplementary Table 3](#)) together with *in-situ* LA-MC-ICP-MS Lu-Hf ([Supplementary Table](#)
711 [4](#)) isotope data for zircons from the adjoining granitoids. Our new Nd isotope data for the
712 greenstone assemblages of the Holenarsipur greenstone belt presented in [Supplementary Table](#)
713 [5](#). Published data of the stratigraphically equivalent greenstone belts and their adjoining
714 granitoids also in the cratonic core have been compiled and also presented in [Supplementary](#)
715 [Tables 1 -5](#) ([Bhaskar Rao et al., 1992, 2008](#); [Monrad, 1983](#); [Peucat et al., 1993](#); [Chardon, 1997](#);
716 [Guitreau et al., 2017](#); [Jayananda et al., 2008, 2015, 2016, 2019](#); [Naqvi et al., 2009](#); [Tushipokla](#)
717 [and Jayananda, 2013](#); [Maya et al., 2017](#); [Mukherjee et al., 2012](#); [Ranjan et al., 2020](#); [Patra et](#)
718 [al., 2020](#); [Panicker et al., 2021](#)).

719 The least altered samples were selected for elemental and Nd isotope analysis. Rocks
720 were chipped into smaller pieces and powdered using a jaw-crusher (tungsten carbide jaws)
721 and agate mill to obtain fine powder. A total of 65 samples (51 greenstone volcanic rocks and
722 14 granitoid samples) were selected for major and trace element measurements. Major elements
723 were analyzed by automated XRF (Pan Analytical Axios Model PW4400/40). The precision
724 for major elements is as follows: 1% for SiO₂, 1.5-3% for Al₂O₃, 2-3% for Fe₂O₃, 10% for
725 MnO, 1-3% for MgO, 2-5% for CaO, 1.5-3% for Na₂O, 2.5% for K₂O, 2-5% for TiO₂, and 5%
726 for P₂O₅. Trace elements for granitoid samples were analysed at National Geophysical
727 Research (NGRI) Hyderabad using HR-ICPMS Nu instruments AttoM. Granitoid samples
728 were dissolved using Microwave acid digestion method (Model MARS-5, CEM corporation,
729 USA) at CSIR-NGRI (Hyderabad) which involves dissolution of 0.05g finely powdered (~200
730 ASTM mesh) sample in a PFA lined vessel using electronic grade HF and analytical grade
731 HNO₃ in 7:3 proportion. The operating parameters of the microwave digestion system are

732 optimally chosen to ramp and hold the processing time to get clear solutions. Rh-103 is used
733 as an internal standard and 0.02% solution of matrix matched geological reference material is
734 used as calibration standard. The data obtained for trace elements generally have a precision
735 better than 5% (RSD) and comparable accuracy. Trace element concentrations for granitoids
736 were determined by using Perkin Elmer SCIEX ELAN DRC II at CSIR-NGRI following the
737 procedures described by [Balaram and Gnanaswar Rao \(2003\)](#).

738 Trace element concentrations for the Holenarsipur volcanic rocks were analyzed at
739 Department of Geosciences, National Taiwan University by quadrupole inductively coupled
740 plasma–mass spectrometry (ICP-MS). About 50 mg sample powders were dissolved in an
741 equal mixture of sub-boiling distilled HF and HNO₃ in a Teflon high-pressure bomb on a hot
742 plate for 48 hours. The dissolved samples were then evaporated to dryness, refluxed with 6 N
743 HNO₃, and reheated close dryness. Then samples were re-dissolved in 2 mL of 3N HNO₃ in
744 high-pressure bombs for an additional 24 h to ensure complete dissolution of refractory
745 minerals. After dissolution, the samples were diluted with Milli-Q water (18.2 mega-ohm) to a
746 final dilution factor of 2000.

747

748 *5.2 Whole-rock Sm-Nd isotopes*

749 Sm-Nd isotope analysis were performed at JAMSTEC (Japan) and Institute of Earth Sciences
750 and Academia Sinica (Taipei). At JAMSTEC, chemical separation procedures for Sm-Nd
751 purification were done using an AG50W-X8 cation ion exchange resin (Bio Rad, California,
752 USA), along with Sr-spec and Ln-spec resins (Eichrom Tec. Inc., Illinois, USA). After the
753 initial separation, using 1 mL of AG50W-X8 resin, Nd was purified using 0.05 mL of Sr-spec
754 resin, 0.3 mL of Ln-spec resin, and 1 mL of Ln-spec resin, respectively. Parts of the column
755 separation procedures were conducted with a fully automated open-column chemical
756 separation system named COLUMNSPIDER ([Miyazaki et al., 2018](#)), developed by JAMSTEC

757 and HOYUTEC Co., Ltd. (Kawagoe, Japan). The total procedural blanks for Nd, and Hf were
758 <69, 8 pg. The Nd isotope ratios were measured with a Triton thermal ionization mass
759 spectrometer (TIMS; Thermo-Finnigan, Bremen, Germany) at JAMSTEC. The measured Nd
760 isotope ratios were normalized to a $^{146}\text{Nd}/^{144}\text{Nd}$ value of 0.7219 to correct for mass
761 fractionation. The mean value obtained by repeated analyses of the JNdi-1 standard was
762 0.512097 ± 0.000009 (2SD, n = 14) which is consistent with the consensus value of 0.512115
763 ± 0.000007 (Tanaka et al., 2000).

764 The analytical procedure at the Institute of Earth Sciences, Academia Sinica (Taipei)
765 involved dissolution of approximately 100 mg of whole-rock powder using a mixture of HF-
766 $\text{HNO}_3\text{-HClO}_4$ in a Teflon beaker at 145°C for seven days. Sample dissolution procedures were
767 repeated to ensure the total dissolution of the refractory phases in the samples. REEs were
768 separated using polyethylene columns with a 5 ml resin bed of AG 50W-X8, 200 to 400 mesh.
769 Neodymium was separated from other REEs using polyethylene columns with a Ln resin as a
770 cation exchange medium. Nd was loaded with H_3PO_4 on a Rhenium double filament. The
771 $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. $^{143}\text{Nd}/^{144}\text{Nd}$ isotope ratios were
772 measured using a Finnigan Triton TIMS at the Institute of Earth Sciences, Academia Sinica,
773 Taipei. JMC Nd standard yielded a $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.511816 ± 12 (2s, n = 101) consistent with
774 the consensus value of JMC standard 0.511812 ± 7 (Jahn et al., 2009). The Chondritic Uniform
775 Reservoir (CHUR) composition used to calculate ϵ_{Nd} values is $^{143}\text{Nd}/^{144}\text{Nd} = 0.512630$ and
776 $^{147}\text{Sm}/^{144}\text{Nd} = 0.1960$ (Bouvier et al., 2008), the ^{147}Sm decay constant is 0.00654 Ga^{-1}
777 (Begemann et al., 2001). Depleted mantle (DM) values used in the calculation of T_{DM} model
778 ages are $^{143}\text{Nd}/^{144}\text{Nd} = 0.51315$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.2137$ (De Paolo, 1981).

779

780 5.3 Zircon U-Pb geochronology by LA-ICP-MS

781 Zircon U–Pb geochronology by LA-ICP-MS was performed at Laboratoire Magmas et Volcans
782 (LMV), Université Clermont Auvergne, Clermont-Ferrand (France) using a Resonetics
783 Resolution M-50 laser-ablation system coupled to a Thermo Element XR. Operating conditions
784 are presented in the [Supplementary Table 6](#). We have handpicked euhedral, clear zircon
785 crystals without visible inclusions as much as possible and mounted them in epoxy. These
786 zircon mounts were further polished and imaged by cathodoluminescence (CL) and/or back-
787 scattered-electrons (BSE) using a Jeol JSM-5910LV scanning electron microscope at LMV.
788 CL/BSE images were acquired using an acceleration voltage of 15 kV. Internal zircon texture
789 was used to select crystals with least radiation damage for U–Pb dating and target most pristine-
790 looking zones (relatively bright zones with fine oscillatory zoning in CL; [Corfu et al., 2003](#)).
791 LA-ICP-MS analyses were carried out using the zircon standard 91500 (1065 Ma: [Wiedenbeck
792 et al. 1995](#)) for external normalization and AS3 (1099 Ma: [Paces & Miller 1993](#)) and Plešovice
793 (337 Ma: [Sláma et al. 2008](#)) for data quality control. Zircon U-Pb data were processed using
794 the GLITTER software ([Van Achterberg et al. 2001](#)), age calculations were done using a
795 present-day $^{238}\text{U}/^{235}\text{U}$ of 137.818 ([Hiess et al., 2012](#)) and decay constants of $1.55125 \cdot 10^{-10}$,
796 $9.8485 \cdot 10^{-10}$, and $4.9475 \cdot 10^{-11} \text{ year}^{-1}$ for ^{238}U , ^{235}U and ^{232}Th , respectively ([Jaffey et al., 1971](#);
797 [LeRoux and Glendenin, 1963](#)). Finally, data were plotted using Isoplot version 3.75 by [Ludwig
798 \(2008\)](#). Zircon reference materials AS3 and Plešovice analyzed during our LA-ICP-MS
799 sessions returned Concordia ages of $1098.2 \pm 2.6 \text{ Ma}$ (2SE, MSWD = 0.68) and 337.6 ± 1.1
800 Ma (2SE, MSWD = 0.70), respectively, which are in excellent agreement with ages provided
801 in [Paces and Miller \(1993\)](#) (1099 Ma) and in [Sláma et al. 2008](#) (337 Ma).

802

803

804 *5.4 Zircon Lu-Hf isotope geochemistry by LA-MC-ICPMS*

805 Lu-Hf isotope measurements were conducted at LMV using a Resonetics Resolution M-50
806 laser-ablation system coupled to a Thermo Neptune Plus MC-ICP-MS (LA-MC-ICP-MS; [see](#)
807 [Supplementary Table 6](#)). Methods and operating conditions are presented in further detail in
808 [Guitreau et al. \(2017, 2019\)](#). Isotopic ratios of $^{173}\text{Yb}/^{171}\text{Yb}$ normalized to the value of 1.132685
809 ([Chu et al., 2002](#)) and $^{179}\text{Hf}/^{177}\text{Hf}$ normalized to the value of 0.7325 ([Stevenson and Patchett](#)
810 [1990](#)) were used to determine Yb and Hf instrumental mass biases. Lu fractionation was
811 assumed to follow that of Yb ([Fisher et al., 2014](#)). Values of 0.79618 for $^{176}\text{Yb}/^{173}\text{Yb}$ ([Chu et](#)
812 [al., 2002](#)) and 0.02655 for $^{176}\text{Lu}/^{175}\text{Lu}$ ([Fisher et al., 2014](#)) were used to correct data for the
813 isobaric interferences of ^{176}Yb and ^{176}Lu on ^{176}Hf . Natural zircon 91500 ([Blichert-Toft, 2008](#))
814 and Mud Tank ([Woodhead and Hergt, 2005](#)) together with the MUN artificial zircon reference
815 materials ([Fisher et al., 2011](#)) were measured multiple times during LA-MC-ICP-MS analyses
816 in order to monitor instrumental fractionation and check data quality. Lu-Hf isotope analyses
817 were done on top of zones with coherent U-Pb systematics that were identified during the U-
818 Pb isotope analyses in order to acquire linked information. Time-resolved signals were
819 carefully monitored throughout each and every analysis to minimize issue of domain mixing
820 and grain heterogeneities. We have used the ^{176}Lu decay constant value of [Scherer et al. \(2011\)](#)
821 and [Söderlund et al. \(2004\)](#), which is $1.867 \cdot 10^{-11} \text{ year}^{-1}$ to calculate initial Hf isotopic
822 compositions. Uncertainties associated with the radiogenic-ingrowth correction were
823 propagated using the algorithms of [Ickert \(2013\)](#). ϵ_{Hf} values were calculated using CHUR
824 values provided in [Iizuka et al. \(2015\)](#). Results on standards and unknowns are available in
825 supplementary Table 4. The synthetic MUN-0 zircon, which contains no measurable quantities
826 of REE, gave an instrumental-fractionation-corrected $^{176}\text{Hf}/^{177}\text{Hf}$ slightly different from that
827 published in ([Fisher et al., 2011](#)) which was used to correct all other zircon reference materials
828 by external normalization. All data were acquired over two separate measurement sessions. For
829 the first session, MUN-1, MUN3, 91500, and Mud Tank gave respective present-day

830 $^{176}\text{Hf}/^{177}\text{Hf}$ of 0.282124 ± 0.000029 , 0.282127 ± 0.000034 , 0.282315 ± 0.000057 , and 0.282511
831 ± 0.000023 (Supplementary Table 4), which are consistent with published values of 0.282135
832 ± 0.000007 for MUN zircons (Fisher et al., 2011), 0.282308 ± 0.000006 for 91500 (Blichert-
833 Toft, 2008), and 0.282507 ± 0.000006 for Mud Tank (Woodhead and Hergt, 2005). Measured
834 $^{176}\text{Lu}/^{177}\text{Hf}$ and $^{176}\text{Yb}/^{177}\text{Hf}$ were also consistent with the range provided in corresponding
835 references. For the second session, MUN-1, MUN-3, MUN-4, 91500, and Mud Tank gave
836 respective present-day $^{176}\text{Hf}/^{177}\text{Hf}$ of 0.282131 ± 0.000027 , 0.282126 ± 0.000027 , $0.282126 \pm$
837 0.000043 , 0.282296 ± 0.000021 , 0.282512 ± 0.000022 , which are also consistent with
838 published values (Supplementary Table 4). The same is true for measured $^{176}\text{Lu}/^{177}\text{Hf}$ and
839 $^{176}\text{Yb}/^{177}\text{Hf}$.

840

841 6. Geochemistry of the greenstone volcanic rocks

842 6.1 Major elements

843 The volcanic rocks of the Holenarsipur greenstone belt are dominantly ultramafic with minor
844 mafic composition. The volcanic rocks of the three blocks in the Holenarsipur greenstone belt
845 show a wide range of elemental compositions. The distinction between the komatiites and
846 komatiitic basalts in the three blocks is based on their respective MgO contents (Arndt and
847 Nisbet, 1982).

848 The volcanic rocks of the southwestern block are komatiitic ($\text{SiO}_2 = 44.3\text{-}47.9$ wt.%
849 and $\text{MgO} = 21.4\text{-}29.7$ wt.%) except two samples that display komatiitic basalt composition
850 ($\text{SiO}_2 = 48.5\text{-}49.1$ wt.%; $\text{MgO} = 15.9\text{-}15.6$ wt.%). Both komatiite and komatiitic basalts exhibit
851 high $\text{CaO}/\text{Al}_2\text{O}_3$ (1.06-1.59) and low $\text{Al}_2\text{O}_3/\text{TiO}_2$ (7.7-14.7) values.

852 Among 20 analyzed samples of volcanic rocks from the northcentral block, five
853 samples show komatiite composition ($\text{SiO}_2 = 45.1\text{-}47.9$ wt.%; $\text{MgO} = 17.9\text{-}23.9$ wt.%) whilst
854 14 samples have komatiitic basalt chemistry ($\text{SiO}_2 = 47.4\text{-}49.3$ wt.%; $\text{MgO} = 16.6\text{-}10.3$ wt.%),

855 one sample show basaltic affinity ($\text{SiO}_2 = 50.6 \text{ wt.}\%$; $\text{MgO} = 9.8 \text{ wt.}\%$) and another one is
856 felsic ($\text{SiO}_2 = 68.6 \text{ wt.}\%$; $\text{MgO} = 1.47 \text{ wt.}\%$; $\text{Na}_2\text{O} = 3.47 \text{ wt.}\%$; $\text{K}_2\text{O} = 1.47 \text{ wt.}\%$). These
857 volcanic rocks (felsic volcanic excluded) display significant variation in $\text{CaO}/\text{Al}_2\text{O}_3$ (1.32-
858 1.09) and $\text{Al}_2\text{O}_3/\text{TiO}_2$ (7.7 – 12.5) values.

859 Among 21 samples analyzed from the eastern block, eighteen samples are volcanic
860 rocks and three samples are gabbro to plagiogranite. Volcanic rocks comprise 12 samples of
861 komatiites ($\text{SiO}_2 = 42.7\text{-}48.0 \text{ wt.}\%$; $\text{MgO} = 35.5\text{-}18.0 \text{ wt.}\%$), 6 samples of komatiitic basalt to
862 high-Mg basalts ($\text{SiO}_2 = 48.5\text{-}50.2 \text{ wt.}\%$; $\text{MgO} = 17.3\text{-}9.4 \text{ wt.}\%$), 2 samples of gabbro ($\text{SiO}_2 =$
863 $48.7\text{-}48.8 \text{ wt.}\%$; $\text{MgO} = 6.73\text{-}6.89 \text{ wt.}\%$) and one plagiogranite sample ($\text{SiO}_2 = 69.7 \text{ wt.}\%$;
864 $\text{MgO} = 0.95 \text{ wt.}\%$; $\text{Na}_2\text{O} = 6.39 \text{ wt.}\%$; $\text{K}_2\text{O} = 0.60 \text{ wt.}\%$). The komatiites show higher
865 $\text{CaO}/\text{Al}_2\text{O}_3$ (1.42-1.04 wt.%), but conversely low $\text{Al}_2\text{O}_3/\text{TiO}_2$ (7.9 to 15.3 except one sample
866 that has higher value of 20.5), compared to komatiitic basalts which exhibit high $\text{CaO}/\text{Al}_2\text{O}_3$
867 (1.04-1.11) and low $\text{Al}_2\text{O}_3/\text{TiO}_2$ (7.92 to 9.42).

868 Published data on volcanic assemblages from the other stratigraphically equivalent
869 greenstone belts in the cratonic core ([Subba Rao and Naqvi, 1999](#); [Jayananda et al., 2008, 2016](#);
870 [Mukherjee et al., 2012](#); [Tushipokla and Jayananda, 2013](#); [Maya et al., 2017](#); [Patra et al., 2020](#))
871 reveal voluminous komatiite and komatiitic basalts displaying large variation in major and
872 trace element compositions. Among the greenstone volcanic rocks, the Nuggihalli samples
873 display a narrow range of $\text{SiO}_2 = 41.3\text{-}42.5 \text{ wt.}\%$ with $\text{MgO} = 33.0\text{-}35.0 \text{ wt.}\%$ but significant
874 variation in $\text{CaO}/\text{Al}_2\text{O}_3$ (1.19-0.61) and $\text{Al}_2\text{O}_3/\text{TiO}_2$ (9.8-21.0) whilst the Kalyadi samples
875 exhibit large variation in SiO_2 (38.9-50.3 wt.%) and MgO (22.9-32.97 wt.%) as well as
876 $\text{CaO}/\text{Al}_2\text{O}_3$ (1.32-0.84) and $\text{Al}_2\text{O}_3/\text{TiO}_2$ (12.38-23.48). The J.C. Pura volcanic rocks also
877 display significant variation in SiO_2 (42.12-50.63 wt.%; but most samples show less than 46.0
878 wt.% SiO_2) and MgO (10.8-42.20 wt.%; most samples show >30 wt.% MgO) as well as
879 $\text{CaO}/\text{Al}_2\text{O}_3$ (2.45-0.70) and $\text{Al}_2\text{O}_3/\text{TiO}_2$ (6.2-22.5). On the contrary, the Ghattihosahalli

880 samples are characterized by a narrow range of SiO₂ (42.0-42.4 wt.%; MgO (39.2- 41.3 wt.%),
881 CaO/Al₂O₃ (1.34-1.32) and Al₂O₃/TiO₂ (5.0). The Banasandra komatiites also display a narrow
882 compositional range of SiO₂ (41.5-46.8 wt.%), high MgO (30.4-42.4 wt.%) and CaO/Al₂O₃
883 (0.64-1.25) and Al₂O₃/TiO₂ (6.75-18.48). The Nagamangala volcanic rocks also exhibit a
884 narrow range of compositions (SiO₂ = 40.5-44.0 wt.%; MgO = 28.8-35.8 wt.%), CaO/Al₂O₃
885 (1.18-0.67) and Al₂O₃/TiO₂ (11.37-19.96).

886 In summary, the volcanic assemblages from the three blocks of the Holenarsipur
887 greenstone belt show variations in major element characteristics wherein the southwestern
888 block shows komatiite with minor komatiitic basalt whilst the northcentral block reveals
889 abundant komatiitic basalts with komatiite, and the eastern block contains komatiite, minor
890 basalt intruded by sheeted dykes, gabbro and plagiogranite. The volcanic sequences from other
891 greenstone belts are dominantly komatiite to komatiitic basalt in composition with significant
892 variations in major element contents. The majority of studied komatiite and komatiitic basalts
893 flows show characteristics similar to the Al-depleted Barberton komatiites and komatiitic
894 basalts with CaO/Al₂O₃ >1 and Al₂O₃/TiO₂ <18 whilst other samples show Al-undepleted
895 [CaO/Al₂O₃ <1.0; Al₂O₃/TiO₂ >18.0] character of Munro-type (Arndt, 2003).

896 Major element data for volcanic rocks of the Holenarsipur greenstone belt together with
897 volcanic rocks of the Nuggihalli, Kalyadi, J.C. Pura, Ghattihosahalli, Banasandra and
898 Nagamangala greenstone belts projected on the triangular plots of Al₂O₃-Fe₂O₃+TiO₂-MgO
899 (after Jensen, 1976 modified by Viljoen et al., 1982) and CaO-MgO-Al₂O₃ (Viljoen et al., 1982)
900 reveal komatiite to komatiitic-basalt composition except few samples extending into the
901 tholeiitic field (Fig.7.1a, b). On the Al₂O₃ versus FeO/(FeO+MgO) versus binary plot (after
902 Viljoen et al., 1982) the volcanic rocks of the southwestern block of the Holenarsipur belt plot
903 on the Barberton komatiite field whilst majority of the samples from the northcentral block fall
904 within the Barberton komatiitic basalt field except few samples (Fig.7.1c). The eastern block

905 samples together with samples from other greenstone belts plot on the Barberton komatiite
906 field and extending into the komatiitic basalt field. About 70% of the studied volcanic rocks
907 show characteristics similar to the Al-depleted Barberton komatiites and komatiitic basalts
908 whilst remaining samples show Al-undepleted character (**Fig.7.1c**) of Munro-type (Arndt,
909 2003).

910 On Harker binary plots, major element oxides plotted against MgO define moderate to
911 strong correlations for greenstone volcanic rocks of the three blocks of the Holenarsipur belt
912 together with volcanic rocks of other equivalent greenstone belts (**Fig.7.1d**) which suggest an
913 olivine ± orthopyroxene control. However, some minor differences in correlations and/or
914 compositions can be seen locally (e.g., TiO₂ vs. MgO in J.C Pura).

915

916 *6.2 Trace elements*

917 The volcanic rocks from the Holenarsipur and other equivalent greenstones from the cratonic
918 core show a wide range of trace element concentrations. Among the Holenarsipur volcanic
919 rocks, komatiites and komatiitic basalts of the southwestern block exhibit high Ni (900-1983
920 ppm except for two samples with 368 and 492 ppm) and Cr (1011-3005 ppm) contents whereas
921 the northcentral block komatiitic basalts and komatiites show low to high Ni (381-1204 ppm)
922 and Cr (941-2167 ppm) concentrations. On the contrary, the eastern block komatiites display
923 moderate to very high Ni contents (796 to 2786 ppm) except for two komatiitic basalts that
924 have lower values of 399 and 486 ppm.

925 The komatiites of the Nuggihalli belt are characterized by high Ni (1544-1960 ppm)
926 and Cr (2702-4400 ppm) concentrations, whilst the Kalyadi komatiites show moderate to high
927 Ni (3145-3177 ppm) and Cr (1969-3597 ppm) contents. In contrast, the J.C. Pura volcanic
928 rocks display large variations in Ni content (600-3064 ppm with exception of a few basaltic
929 samples that have less than 500 ppm), and Cr (800-5649 ppm with a few samples less than 700

930 ppm). The Ghattihosahalli komatiites exhibit the highest Ni contents (3145-3147 ppm) but
931 conversely low Cr (1208-1338 ppm) among ultramafic volcanic rocks of the region, whilst the
932 Banasandra komatiites show high Ni concentrations (1313-2619 ppm) with a large variation in
933 Cr (1614-2494 ppm). The Nagamangala komatiites are characterized by high Ni (1121-1590
934 ppm) with a wide range of Cr contents (1294-3860 ppm).

935 When projected against MgO in Harker's binary diagrams, petrogenetically significant
936 trace elements show strong to poorly fractionated trends of the studied volcanic rocks
937 (Fig.7.1e). The greenstone volcanic rocks in the cratonic core show significant variations in
938 the REE and other incompatible element abundances. The studied komatiites and komatiitic
939 basalts from the three blocks of the Holenarsipur belt exhibit large variations in REE contents.
940 Samples from the SW block form two groups (low- and high-REE; Fig.7.2a, c). The low REE
941 samples (Σ REE = 4.4-19.2 ppm; MgO = 25.1-29.7 wt.%) exhibit flat REE patterns with
942 variable $(\text{Gd}/\text{Yb})_{\text{N}}$ values of 0.53-1.86 (see fig.7.2a). On primitive-mantle normalized multi-
943 element diagrams (Sun and McDonough, 1989), these samples display either positive or no Zr-
944 Ti anomalies without any Hf anomalies whilst all samples show slight positive Y anomalies
945 (Fig.7.2b). The high REE samples (Σ REE = 20.6-66.7 ppm; MgO = 15.6-29.5) show
946 fractionated REE with $(\text{Gd}/\text{Yb})_{\text{N}}$ of 1.22-2.29 (see fig.7.2c) with two samples displaying
947 negative Eu anomalies. On the contrary, the high Σ REE komatiites show either slightly positive
948 or no significant Nb anomalies associated with negative Zr, Hf, Ti and Y anomalies on
949 primitive-mantle normalized multi-element diagrams (Fig.7.2d). Two samples exhibit negative
950 Nb anomalies coupled with mild positive Zr, Hf and Y anomalies on multi-element diagrams
951 (see fig.7.2d).

952 Komatiites from the northcentral block (MgO = 17.9-23.9 wt.%) with low to high Σ REE
953 (14.6 -59.1 ppm) show flat to moderately fractionated REE patterns with three samples having
954 negative Eu anomalies (Fig.7.2e). On the other hand, komatiitic basalts contain moderate to

955 high Σ REE (31.2-138.0 ppm; MgO = 10.3-16.6 wt.%) and display flat to moderately
956 fractionated REE patterns (**Fig.7.2g**) with one sample displaying a negative Eu anomaly. On
957 primitive-mantle normalized multi-element diagrams, both komatiites (**Fig.7.2f**) and komatiitic
958 basalts (**Fig.7.2h**) exhibit negative Nb anomalies with slight positive to negative Zr, Hf, Ti, Y
959 anomalies. The felsic volcanic sample shows [high SiO₂ = 68.65 wt.%, Al₂O₃ = 14.94 wt.%,
960 MgO = 1.29 wt.%, and Na₂O = 3.17 wt.%] high total REE (392 ppm) content and poorly
961 fractionated REE patterns [(La/Yb)_N = 2.47], as well as a slightly negative Eu anomaly (**see**
962 **fig.7.2g**). The samples display either negative or no significant Nb, Zr, Hf, Ti and Y anomalies
963 on primitive-mantle normalized multi-element diagram.

964 The eastern block volcanic rocks are mainly komatiitic with a few samples displaying
965 basaltic compositions, and their REE contents reveal two groups (i.e., low- and high-REE).
966 The low Σ REE group (4.4-18.4 ppm with MgO = 22.8-30.9 wt.%) is characterized by flat and
967 poorly to moderately fractionated REE patterns [(Gd/Yb)_N = 0.41-1.38] with two samples
968 showing negative Eu anomalies (**Fig.7.2i**). On PM-normalized multi-element diagrams, no
969 significant Nb, Zr and Hf anomalies but positive Ti anomalies and variable Y anomalies (either
970 positive, negative or no anomaly; **Fig.7.2j**). The high Σ REE (48.4-294.8 ppm) group (MgO =
971 17.0-30.4 wt.%) exhibits fractionated REE patterns (Gd/Yb)_N = 1.16-1.96; **Fig.7.2k**) coupled
972 with positive Nb anomalies and negative Zr, Hf and Ti anomalies (**Fig.7.2l**) on multi-element
973 diagrams. Two samples of gabbro intruded into the komatiite to basaltic volcanic rocks in the
974 eastern block show low total REE contents (34.4-44.3 ppm), LREE depletion and poorly
975 fractionated REE patterns with (Gd/Yb)_N values ranging from 0.92 to 1.05 (**Fig.7.2i**). A
976 plagiogranite unit injected into the volcanic sequence and gabbro shows low total REE content
977 (41.7 ppm) with poorly fractionated REE pattern and (La/Yb)_N value of 5.66 (**see fig.7.2i**).

978 Komatiites from the Nuggihalli greenstone belt exhibit sub-chondritic to low total REE
979 concentrations (Σ REE = 4.2-30.9 ppm) and display poor to moderately fractionated REE

980 patterns [(Gd/Yb)_N = 0.43-1.24] with one sample showing negative Eu anomaly (**Fig.7.3a**). On
981 PM normalized multi-element plot, two samples show positive Zr, Ti and Y anomalies whilst
982 one sample shows negative Zr, Ti and Y anomalies (**Fig.7.3b**). Komatiites of the Kalyadi
983 greenstone belt also display sub-chondritic to moderate total REE contents (Σ REE = 2.7-52.0
984 ppm) with poorly to moderately fractionated REE patterns [(Gd/Yb)_N = 0.62-1.48] except for
985 one sample that shows negative Ce anomaly (**Fig.7.3a**). On the PM normalized multi-element
986 diagram, three samples exhibit positive Zr and Ti anomalies, but two samples show positive Y
987 anomalies and one sample show negative Nb, Zr and Ti anomalies (**see Fig.7.3b**).

988 The komatiite and komatiitic basalts of the J.C Pura greenstone belt exhibit large
989 variations in REE abundances (**Jayananda et al., 2008, 2016**) wherein samples of komatiite
990 flows (<35 wt.% MgO) exhibit low to moderate REE (Σ REE = 14.1-45.9 ppm) whilst the
991 cumulate layers (>35 upto 43 wt.%) contain very low to sub-chondritic REE concentrations
992 (3.6-12.1 ppm). The samples of komatiite flows show uniform patterns with flat to moderately
993 fractionated REE patterns [(Gd/Yb)_N = 0.98-1.85] except few samples displaying negative Eu
994 anomalies (**Fig.7.4a**). On PM normalized multi-element diagrams, they exhibit strong negative
995 Zr, Hf, Ti and Y anomalies with positive Nb anomalies (**Fig.7.4b**). The cumulate layers show
996 flat to moderately fractionated REE patterns with [(Gd/Yb)_N = 0.65-1.66] with two samples
997 showing either positive or negative Eu anomalies (**Fig.7.4c**). On PM normalized multi-element
998 plots, these cumulate layers show slight positive Nb and Y anomalies but significant negative
999 Hf and positive Ti anomalies with a few samples showing no or very faint anomalies
1000 (**Fig.7.4d**). The J.C. Pura komatiitic basalts display poor to moderately fractionated REE
1001 patterns [(Gd/Yb)_N = 0.66-1.48] whilst on the PM normalised multi-element diagram they
1002 exhibit negative Nb, Zr, Hf, and Y anomalies (**Fig.7.4e,f**).

1003 The Ghattihosahalli komatiites have very low REE concentrations (Σ REE = 9.76-9.81
1004 ppm) and show fractionated REE patterns with (Gd/Yb)_N = 2.26-2.52 (**Fig.7.5a**). On PM

1005 normalized multi-element diagrams, they show slight negative Nb and Th anomalies with
1006 strong negative Hf anomalies (**Fig.7.5b**). The Banasandra komatiites exhibit sub-chondritic to
1007 low total REE contents (3.5-12.2 ppm) and show poorly to moderately fractionated REE
1008 patterns [(Gd/Yb)_N = 0.99-1.90)] (**Fig.7.5c**). On PM normalized multi-element diagrams, these
1009 komatiites display positive Nb and Ti anomalies associated with negative Hf anomalies
1010 (**Fig.7.5d**). The Nagamangala komatiites and komatiitic basalts contain low to moderate total
1011 REE contents (9.2-51.0 ppm), and exhibit poorly to moderately fractionated REE [(Gd/Yb)_N =
1012 0.59-2.57)] (**Fig.7.5e**). On PM normalised multi-element plot, these samples show slight
1013 negative Nb but strong negative Hf anomalies (**Fig.7.5f**).

1014 When projected on (Gd/Yb)_N versus Al₂O₃/TiO₂ and CaO/Al₂O₃ plots, data for
1015 komatiite and komatiitic basalts from all three blocks of the Holenarsipur greenstone belt,
1016 together with volcanic rocks of other greenstone belts from the cratonic core form an ellipse
1017 that overlaps fields corresponding to Barberton and Abitibi-type komatiites. About 70% of data
1018 fall in the field defined by Barberton komatiites and the remaining samples plot in the field of
1019 Abitibi-type komatiites (**Fig.7.6a, b**).

1020

1021 6.3 Sm-Nd isotope geochemistry

1022 Our new Sm-Nd isotope data for the Holenarsipur greenstone belt along with the published
1023 data for other greenstone belts (Jayananda et al., 2008; Mukherjee et al., 2012; Maya et al.,
1024 2017; Patra et al., 2020) are presented in the [Supplementary Table 5](#). The komatiite and
1025 komatiitic basalts from all the greenstone belts in the cratonic core show significant variations
1026 in ¹⁴⁷Sm/¹⁴⁴Nd and ¹⁴³Nd/¹⁴⁴Nd values. Among the Holenarsipur volcanic rocks, komatiite and
1027 komatiitic basalts from the southwestern block displays a well-defined positive correlation (R²
1028 = 0.99) between present-day ¹⁴⁷Sm/¹⁴⁴Nd and ¹⁴³Nd/¹⁴⁴Nd that show a range of respectively,
1029 0.1137-0.2105 and 0.510971-513037. The higher the Nd and Sm concentrations, the lower the

1030 $^{147}\text{Sm}/^{144}\text{Nd}$. The slope of this correlation translates into a Sm-Nd regression age of 3118 ± 168
1031 Ma that gives an initial $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.50870 ± 0.00021 , which translates into an initial ϵ_{Nd}
1032 of $+2.1 \pm 4.1$ at 3118 Ma.

1033 Volcanic rocks from the eastern block exhibit positively correlated ($R^2 = 0.85$)
1034 $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ (present-day) ranging from, respectively, 0.1318-0.2211 and
1035 0.511482-0.513980. The higher Nd and Sm concentrations, the lower $^{147}\text{Sm}/^{144}\text{Nd}$. Sm-Nd
1036 regression for Eastern block volcanic rocks give an imprecise age of 3043 ± 428 Ma and an
1037 initial $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.50876 ± 0.00037 , which translates into an initial ϵ_{Nd} of $+1.4 \pm 7.3$ at
1038 3043 Ma.

1039 Volcanic rocks from the northcentral block present weakly correlated ($R^2 = 0.62$)
1040 $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ (present-day) that range from, respectively, 0.1695-0.2139 and
1041 0.512127-0.513119. If the Sm-Nd correlation is converted into a regression age, it gives 2578
1042 ± 763 Ma associated with an initial $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.5094 ± 0.0013 , which translates into an
1043 initial a ϵ_{Nd} of $+2.0 \pm 2.3$ at 2578 Ma that is comparable with the Sm-Nd whole rock isochron
1044 age 2620 ± 55 Ma documented by [Drury et al \(1987\)](#). However, our $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$
1045 data for komatiites and komatiitic basalts from all the three blocks of the Holenarsipur belt
1046 together with published data for komatiites ([Jayananda et al., 2008](#)) from other stratigraphically
1047 equivalent greenstone belts in the cratonic core yield a regression age of 3356 ± 68 Ma for the
1048 komatiite volcanism in the Holenarsipur belt which is associated with an initial $^{143}\text{Nd}/^{144}\text{Nd}$ of
1049 0.50843 ± 0.00027 that translates into an initial ϵ_{Nd} of $+2.9 \pm 5.3$ at 3356 Ma. Our data exhibit
1050 some consistency with published Sm-Nd isotope data for WDC komatiites ([Jayananda et al.,](#)
1051 [2008](#)) because they follow the main $^{147}\text{Sm}/^{144}\text{Nd}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ correlation that provided an age
1052 of 3352 ± 110 Ma.

1053 The volcanic rocks of other greenstone belts in the cratonic core display large variations
1054 in $^{147}\text{Sm}/^{144}\text{Nd}$, $^{143}\text{Nd}/^{144}\text{Nd}$ and $\epsilon_{\text{Nd(T)}}$. The Nuggihalli and Kalyadi volcanic rocks exhibit high

1055 $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ values with $\epsilon_{\text{Nd}(T)}$ signatures (Jayananda et al 2008; Mukherjee et
1056 al., 2012; Patra et al, 2020) ranging from +1.3 to +5.4 except for three samples that have
1057 negative signatures (i.e., -1.4 to -3.6). These Sm-Nd data suggest their origin from variably
1058 depleted mantle reservoirs with possible traces of crustal contamination or influence of
1059 secondary processes. The moderate to high $^{147}\text{Sm}/^{144}\text{Nd}$ (0.1851 – 0.2194) and $^{143}\text{Nd}/^{144}\text{Nd}$
1060 (0.511569 to 0.513203), and $\epsilon_{\text{Nd}(T)}$ of +1.2 to +2.6 for the J.C. Pura komatiites can be attributed
1061 to their origin from moderately depleted mantle reservoirs. The Ghattihosahalli samples display
1062 relatively narrow range of $^{147}\text{Sm}/^{144}\text{Nd}$ (0.1493 – 0.1686) and $^{143}\text{Nd}/^{144}\text{Nd}$ (0.511859 to
1063 0.512159) values and $\epsilon_{\text{Nd}(T)}$ from +2.8 to +5.3 indicating depleted mantle reservoirs. In contrast,
1064 Banasandra komatiites show large variations in $^{147}\text{Sm}/^{144}\text{Nd}$ (0.1224 – 0.2431) and $^{143}\text{Nd}/^{144}\text{Nd}$
1065 (0.511174 to 0.514037) with $\epsilon_{\text{Nd}(T)}$ from +1.1 to +5.1 suggesting their origin from variably
1066 depleted mantle reservoirs.

1067

1068 *6.4 Timing of greenstone volcanism (ca. 3400-3200 Ma)*

1069 The time frame of greenstone volcanism in the cratonic core is poorly constrained as precise
1070 and accurate geochronological data are rather scanty. Most direct geochronological
1071 information pertaining to greenstone volcanism have been obtained using whole-rock Sm-Nd
1072 isotope system or through U-Pb dating of zircons from interlayered felsic volcanic units (e.g.
1073 Peucat et al., 1995). Dating komatiite assemblages is challenging due to their unusually low
1074 Nd contents coupled with the effects of metamorphism and hydrothermal alteration processes
1075 which affected the Sm-Nd isotope system. Drury et al. (1987) presented a Sm-Nd whole-rock
1076 isochron age of 2620 ± 55 Ma for the mafic volcanic rocks from the “so called” younger
1077 Dharwar Supergroup of the Holenarsipur belt (Swami Nath and Ramakrishnan, 1981)
1078 corresponding to northcentral block of the present study. In contrast, Peucat et al (1995)
1079 obtained U-Pb zircon age of 3298 ± 7 Ma for felsic volcanic flow from the same younger

1080 Dharwar Supergroup (Swami Nath and Ramakrishnan, 1981) and concluded that the entire
1081 Holenarsipur belt belongs to the older Sargur Group. Our new U-Pb zircon age of 3305 ± 10
1082 Ma (Fig.7.8m) from the same felsic volcanic flow also confirms the age obtained by Peucat et
1083 al (1995). Further, $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ data for komatiites and komatiitic basalts from
1084 the three blocks of the Holenarsipur belt together with published data for komatiites from other
1085 stratigraphically equivalent greenstone belts from the cratonic core yield an isochron age of
1086 3356 ± 68 Ma. Our data exhibit some consistency with published Sm-Nd isotope data of
1087 komatiites (Jayananda et al., 2008) because they follow the main $^{147}\text{Sm}/^{144}\text{Nd}$ - $^{143}\text{Nd}/^{144}\text{Nd}$
1088 correlation that provided an age of 3352 ± 110 Ma based on 16 samples from the J.C. Pura,
1089 Nuggihalli, Kalyadi, Ghattihosahalli and Banasandra greenstone belts. The J.C. Pura
1090 komatiites alone define a Sm-Nd whole-rock isochron age of 3384 ± 200 Ma (MSWD = 1.8).
1091 Published Sm-Nd data for komatiites from Nuggihalli and adjoining Kalyadi greenstone belt
1092 (Jayananda et al., 2008; Mukherjee et al., 2012; Patra et al., 2020) together define a whole-rock
1093 isochron age of 3207 ± 89 Ma (initial $^{143}\text{Nd}/^{144}\text{Nd} = 0.00012$; MSWD=4.3) which is similar to
1094 the ages obtained for these belts 3284 ± 310 Ma (MSWD 3.7; this work and Jayananda et al
1095 2008) as well as recent Sm-Nd and Lu-Hf isochron ages of komatiites and komatiitic basalts
1096 from the north-eastern part of Bababudan basin (Ravindran et al., 2021). Sm-Nd isotope data
1097 for peridotite-pyroxenite-anorthosite-gabbro from the Nuggihalli belt yielded a whole-rock
1098 isochron age of 3125 ± 120 Ma (Mukherjee et al., 2012). The Banasandra komatiites define a
1099 Sm-Nd whole-rock isochron age of ca. 3150 Ma (Maya et al., 2017). Although other workers
1100 (Mukherjee et al., 2012; Maya et al., 2017) have obtained Sm-Nd whole-rock isochron ages of
1101 3150 Ma for the Nuggihalli and Banasandra greenstone belts, they are within the limits of Sm-
1102 Nd and Lu-Hf whole-rock isochron results considering their elevated analytical uncertainties.
1103 Our new Sm-Nd isotope data of the komatiites and komatiitic basalts from the three blocks of

1104 the Holenarsipur greenstone belt together with published data define a whole-rock isochron
1105 age of 3356 ± 68 Ma for the widespread komatiite volcanism in the cratonic core.

1106 On the other hand, komatiite and komatiitic basalts from the Nuggihalli, Kalyadi, J.C.
1107 Pura, Ghattihosahalli and Banasandra greenstone belts exhibit wide ranges of $^{147}\text{Sm}/^{144}\text{Nd}$
1108 values which is reflected in their LREE contents (Jayananda et al., 2008; Mukherjee et al., 2012;
1109 Maya et al., 2017; Patra et al., 2020). Twenty-one samples from different greenstone belts
1110 (excluding Holenarsipur) define a whole-rock isochron age of 3330 ± 210 Ma. Whole-rock
1111 isochrons calculated for the komatiite and komatiitic basalts of individual greenstone belts
1112 (Jayananda et al., 2008; Mukherjee et al., 2012; Patra et al., 2020; Maya et al. 2017; this study)
1113 exhibit similar initial Nd isotope compositions which suggest a genetic link. Sargur Group
1114 greenstones could, hence, have been erupted within short-time intervals in which case Sm-Nd
1115 age dispersion would be due to secondary (post-crystallization) processes. Alternatively,
1116 greenstone volcanic assemblages emplaced over a long time period and the range of ~ 210 Ma
1117 indicates an actual time span over which greenstone volcanic rocks formed, hence, essentially
1118 between 3400 and 3200 Ma concurrently with WDC granitoids formation (Peucat et al., 1993;
1119 Jayananda et al., 2015; Guitreau et al., 2017; Ranjan et al., 2020; Bidyananda et al., 2016).

1120

1121 **7. Geochemistry of granitoids (TTG-type gneisses) and diapiric trondhjemites**

1122 *7.1 Zircon U-Pb geochronology*

1123 Our new U-Pb zircon data on granitoids, felsic volcanic flow of the Northcentral block
1124 and a plagiogranite from Eastern block together with published ages on the gneisses adjacent
1125 to the Holenarsipur greenstone belt presented in the [Supplementary Table 3](#). Furthermore, CL
1126 images ([Fig. 7.7a-c](#)) of the studied zircons and U-Pb concordia presented in [Fig. 7.8a-n](#). Our
1127 new zircon age data together with published U-Pb zircon ages of the gneisses and trondhjemite

1128 intrusions of the cratonic core (Peucat et al., 1993; Jayananda et al., 2015, 2019; Ranjan et al.,
1129 2020) are described in the following section.

1130 Zircon crystals from the sample 18HP23 commonly exhibit fine oscillatory zoning (Fig.
1131 7.7c) with various degrees of radiation damage accumulation. Th/U ratios are essentially within
1132 magmatic values (0.01-0.81 with an average of 0.45) and U concentrations are low to very high
1133 (177-1419 ppm with an average of 668 ppm). U-Pb data display various degrees of discordance
1134 but align along a single Discordia line with an upper-intercept represented by concordant
1135 analyses giving a Concordia age of 3430 ± 7 Ma (2 SE; Fig.7.8a) that we interpret as that of
1136 18HP23 crystallization. One concordant analysis returned an age of 3500 ± 12 Ma (2 SE) which
1137 we interpret as an inherited crystal because the core of this zircon looks distinct from the rest
1138 of the crystal which has fine oscillatory zoning.

1139 Most zircon crystals from the sample 18HP24a exhibit fine oscillatory or convolute
1140 zoning (Fig. 7.7c) with evidence for local radiation damage accumulation. Th/U ratios vary
1141 between 0.21 and 1.24 with an average of 0.74, hence, within magmatic values. U
1142 concentrations are low to very high because they vary between 55 and 1497 ppm with an
1143 average of 524 ppm. U-Pb data are variably discordant but align along a single Discordia line
1144 that intercepts the Concordia at an age of 3431 ± 8 Ma (2 SE) given by concordant analyses.
1145 This age is interpreted as that of crystallization of 18HP24a (Fig.7.8b).

1146 Zircon crystals from the sample 18HP10b, which is a mica-rich enclave, provided three
1147 concordant ages of 3468 ± 17 Ma, 3271 ± 11 Ma, and 3166 ± 9 Ma (Fig.7.8c). These zircon
1148 crystals show bright fine oscillatory zoning with darker zones due to radiation damage
1149 accumulation (Fig. 7.7c). Th/U ratios vary between 0.05 and 0.58 with an average of 0.44,
1150 hence, within magmatic values, and U concentrations are very low to high (12-868 ppm;
1151 average 245 ppm).

1152 The sample 18HP05 yielded very few zircon crystals of acceptable quality for U-Pb
1153 dating. We, therefore, report results for four crystals that exhibit oscillatory zoning with oval
1154 resorption patterns underlined by thin bright zones (Fig. 7.7b). Many crystals have
1155 accumulated large degrees of radiation damage. Th/U ratios vary between 0.59 and 0.87, hence,
1156 clearly within the magmatic range and U concentrations are low (83-431 ppm). U-Pb data are
1157 limitedly discordant and align along a single discordia line consistent with the concordia age
1158 of 3341 ± 9 Ma (2 SE; Fig.7.8d) given by the concordant data. This age is interpreted as the
1159 crystallization age of the sample 18HP05.

1160 Zircon crystals from the sample 18HP01a exhibit various internal textures from bright
1161 and fine oscillatory zoning to dark patchy domains (Fig. 7.7a). Only best-looking domains
1162 were analysed and returned Th/U ratios from 0.04 to 1.37, with the majority being around the
1163 average value of 0.55. Uranium concentrations are very low to high (34-713 ppm) with an
1164 average of 221 ppm. U-Pb data are variably discordant and do not align along a single Discordia
1165 line but define a fan shape that converges towards concordant data indicating a Concordia age
1166 of 3319 ± 7 Ma (2 SE; Fig.7.8e). This age is interpreted as that of magmatic crystallization of
1167 the sample 18HP01a.

1168 Zircon crystals from the sample 18HP01c are commonly dark in CL images and have
1169 broad zoning and stubby habit (Fig. 7.7a). Th/U ratios are consistent between crystals from
1170 this population (0.22-0.77) and so are U concentrations (281-757 ppm: average 501 ppm). U-
1171 Pb data are variably discordant but most concordant data record a concordia age of 3311 ± 10
1172 Ma (2 SE; Fig.7.8f), assumed to be that of magmatic crystallization of the sample 18HP01c.

1173 Zircon crystals from the sample 18HP03a have either fine oscillatory zoning or broad
1174 dark zoning (Fig.7.7a) with various evidence for advanced radiation damage accumulation.
1175 Th/U ratios are mostly high and vary between 0.40 and 1.81 (average of 0.94). Uranium
1176 concentrations are essentially low (27-295 ppm). U-Pb data are variably discordant and do not

1177 align along a single discordia but converge towards concordant data that give a concordia age
1178 of 3320 ± 8 Ma (2 SE; **Fig.7.8g**), interpreted as that of magmatic crystallization of the
1179 sample 18HP03a.

1180 Zircon crystals from sample 18HP03b commonly display fine oscillatory zoning
1181 although effects of radiation damage are also visible in some portions of the crystals (**Fig. 7.7a**).
1182 Th/U are commonly alike magmatic values (0.24-0.77) except for 2 crystals (0.05 and 1.12)
1183 and U concentrations are very variable (86-2940 ppm). U-Pb data show great dispersion, but
1184 some concordant data provide a Concordia age of 3320 ± 8 Ma (2 SE; **Fig.7.8h**) that is thought
1185 to be that of igneous crystallization of the sample 18HP03c.

1186 Zircon crystals from the sample 18HP04a are generally dark in CL images but show
1187 fine oscillatory zoning (**Fig. 7.7a, b**). Th/U ratios fall within magmatic values because they
1188 range from 0.43 to 0.91. Uranium concentrations are low to moderate (209-764 ppm; average
1189 = 418 ppm). U-Pb data distribute close to the Concordia curve and most concordant data give
1190 an age of 3304 ± 8 Ma (2 SE; **Fig.7.8i**) which is assumed to be that of crystallization.

1191 Zircon crystals from the sample 18HP04b exhibit fine oscillatory zoning as well as
1192 sector zoning (**Fig. 7.7b**). Their Th/U vary between 0.48 and 1.32 with an average of 395
1193 whereas U concentrations are generally moderate (253-568 ppm). U-Pb data show complex
1194 characteristics in Concordia plot that returned two concordant clusters at 3292 ± 9 Ma (2 SE)
1195 and 3339 ± 8 Ma (**Fig.7.8j**). We interpreted the first age as that of magmatic crystallization and
1196 the second one as zircon inheritance.

1197 Zircon crystals from the sample 18HP07 commonly have simple fine oscillatory zoning
1198 (**Fig. 7.7b**) despite some crystals having obvious signs of advanced radiation damage. Th/U
1199 are essentially within magmatic values (0.07-0.73) and U concentrations are very variable
1200 (179-1185 ppm). U-Pb data essentially align along a single discordia which upper-intercept

1201 correspond to concordant data that provided a concordia age of 3429 ± 9 Ma (2 SE; [Fig.7.8k](#)),
1202 assumed to be that of 18HP07 crystallization.

1203 Zircon crystals from the sample [18HP15a](#) are characterized by fine oscillatory zoning
1204 ([Fig. 7.7c](#)) and Th/U between 0.14 and 0.77, which corresponds to magmatic values. Their U
1205 contents are low to very high because they vary from 216 to 1773 ppm with an average of 575
1206 ppm. With the exception of a few datapoints, U-Pb data align along a single discordia line with
1207 an upper-intercept that coincides with most concordant data at 3429 ± 11 Ma (2 SE; [Fig.7.8l](#)),
1208 most likely indicative of the time of igneous crystallization.

1209 The sample [18HP13](#) yielded very poor-quality zircons except for one that we analysed
1210 multiple times. This crystal revealed a simple oscillatory zoning ([Fig. 7.7c](#)), consistent Th/U
1211 (0.24-0.71) and U concentrations (43-281 ppm). U-Pb data are either concordant or limitedly
1212 discordant all consistent with an age of 3305 ± 10 Ma (2 SE; [Fig.7.8m](#)).

1213 The few zircon crystals that we managed to recover from sample [18HP39](#) show fine oscillatory
1214 zoning ([Fig. 7.7c](#)) and generally simple textures. Their Th/U are magmatic (0.35-0.82) and U
1215 concentrations variable (180-1001 ppm). U-Pb data gave a coherent age of 3347 ± 11 Ma (2
1216 SE) that we interpret as that of crystallization ([Fig.7.8n](#)).

1217 Published U-Pb zircon age data on the granitoids and detrital zircons from the cratonic
1218 core around Holenarsipur-Gorur, southern margin Bababudan basin, the J.C. Pura and Bellur-
1219 Nagamangala regions reveal that their magmatic protoliths formed during 3600-3200 Ma
1220 ([Nutman et al., 1992](#); [Meen et al., 1992](#); [Peucat et al., 1993](#); [Bidyananda et al., 2016](#); [Jayananda](#)
1221 [et al., 2015](#); [Ranjan et al., 2020](#); [Ao et al., 2021](#)). The oldest zircon xenocrysts from
1222 trondhjemitic gneiss in the Hassan-Gorur region indicate 3610 Ma ([Guitreau et al., 2017](#)) whilst
1223 detrital zircons from pelitic sedimentary rocks in the same region provide ages of 3560-3500
1224 Ma ([Nutman et al., 1992](#); [Ranjan et al., 2022](#)). More recent U-Pb zircon ages of the granitoids
1225 (TTGs) reveal 3350-3270 Ma for formation of their magmatic precursors ([Jayananda et al.,](#)

1226 [2015; Ranjan et al., 2020](#)). Northeast of the Holenarsipur greenstone belt, detrital zircons from
1227 quartzite in the J.C Pura greenstone belt and pelites from the western boundary of the
1228 Chitradurga greenstone belt provided U-Pb ages of 3400 - 3200 Ma ([Ramakrishnan et al., 1994](#);
1229 [Hokada et al., 2013](#); [Lancaster et al., 2015](#); [Wang and Santosh, 2019](#)). South of the
1230 Holenarsipur, in the Kabini dam area, older detrital zircons from paragneiss also indicate U-Pb
1231 ages of 3400 - 3200 Ma ([Bidyananda et al., 2016](#)).

1232

1233 *7.2 Major episodes of granitoid crust formation*

1234 U-Pb zircon ages of the present study together with published ages for the granitoids of the
1235 WDC show three major stages of felsic crust formation marking episodic continental growth
1236 during Paleoproterozoic ([Peucat et al., 1993](#); [Jayananda et al., 2015](#); [Guitreau et al., 2017](#); [Ranjan](#)
1237 [et al., 2020](#)).

1238 The granitoids in the southwestern block adjoining to the Holenarsipur greenstone belt
1239 emplaced during three major events. Zircon xenocrysts from the trondhjemite ([Guitreau et al.](#)
1240 [2017](#)) together with ages of zircon xenocrysts from this study indicate U-Pb ages of 3610-3500
1241 Ma corresponding to the earliest event ([Guitreau et al., 2017](#)). U-Pb zircon data for the tonalitic
1242 to granodioritic gneisses record a second episode at 3470-3400 Ma whilst several U-Pb zircon
1243 ages for tonalitic to the granodioritic gneisses indicate a major crust building event during the
1244 time interval 3350-3270 Ma. Our new zircon U-Pb ages for the gneisses adjoining to the
1245 northcentral block provided ages of 3429 ± 9 Ma and 3429 ± 11 Ma, respectively ([see fig.7.8k,](#)
1246 [I](#)) without any identified inherited cores. Recent U-Pb zircon dating of the granitoids close to
1247 our sampled locations ([see fig. 2b](#)) revealed magmatic crystallization events of their precursors
1248 at 3361 ± 25 Ma, 3339 ± 23 Ma and 3302 ± 44 Ma ([Ao et al., 2021](#)). Although stratigraphic
1249 relationships between the greenstone sequences of the northcentral block and surrounding
1250 granitoids (gneisses) remain uncertain due to commonly tectonized contacts, greenstone

1251 volcanism and surrounding granitoid formation may be sub-contemporaneous whereas $3305 \pm$
1252 10 Ma ([see fig.7.8m](#)) pyroclastic rhyolite flows may represent a terminal stage of greenstone
1253 volcanism in the northcentral block.

1254 In the eastern block, zircons from the plagiogranite interlayered with gabbroic rocks in
1255 the northern roadcut yielded a U-Pb age of 3347 ± 11 Ma ([see fig.7.8n](#)) which coincides with
1256 imprecise Sm-Nd whole rock isochron age and of 3345 ± 111 Ma defined for the associated
1257 komatiites and komatiitic basalts. Published U-Pb zircon ages for tonalitic gneiss from the
1258 western margin of the eastern block indicate ages of 3276 ± 5 Ma ([Jayananda et al., 2015](#)) and
1259 3289 ± 6 Ma ([Ranjan et al., 2020](#)). Thus, the eastern granitoids are slightly younger than
1260 adjoining greenstone in which the plagiogranite injected into komatiitic to basaltic volcanic
1261 rocks defined an age of 3347 ± 11 Ma ([see fig.7.8n](#)).

1262 Diapiric trondhjemites intrude into the volcanic assemblages of all the three blocks of
1263 the Holenarsipur greenstone belt and post-date the formation of granitoids. Published U-Pb
1264 zircon and titanite ages together with new data from the present study indicate 3230-3177 Ma
1265 with xenocrysts and inherited cores up to 3350 Ma (present study; [Peucat et al., 1993](#);
1266 [Jayananda et al., 2015](#); [Guitreau et al., 2017](#); [Ranjan et al., 2020](#)).

1267 Published U-Pb zircon ages for the granitoids adjacent to the J.C. Pura greenstone belt
1268 indicate U-Pb zircon ages of 3315-3270 Ma ([Chardon, 1997](#)) whilst Pb-Pb ages of feldspar
1269 separates from gneisses around the North-eastern part of the Holenarsipur belt and the Kalyadi
1270 greenstone belt provide ages 3350 Ma with an inherited component as old as 3800 Ma ([Meen](#)
1271 [et al., 1992](#)). Zircon crystals from granitoids gneisses along the western margin of the
1272 Chitradurga and Nagamangala belts provides U-Pb and Pb-Pb ages ranging from 3315 to 3270
1273 Ma ([Taylor et al., 1984, 1988](#); [Ravindran et al., 2020](#); [Jayananda et al., 2019](#)). On the other
1274 hand, the granitoids in the southern and eastern margin of the Bababudan belt indicate U-Pb

1275 zircon ages of 3350-3200 Ma with inherited crystals as old as 3600-3800 Ma (Peucat et al.,
1276 1993; Jayananda et al., 2015; Bidyananda et al., 2016).

1277 Diapiric trondhjemite plutons intruding the TTG-type granitoids adjoining to the J.C
1278 Pura greenstone belt indicate an age of 3158 ± 9 Ma (Ao et al., 2021) which coincides with a
1279 U-Pb zircon age of 3198 ± 26 Ma obtained from aplitic dykes that intruded the TTG-type
1280 granitoids (Chardon, 1997).

1281

1282 *7.3 Elemental geochemistry*

1283 The granitoids adjacent to the three greenstone units of the Holenarsipur greenstone belt
1284 comprises heterogeneous association of grey gneisses with a TTG affinity. The gneisses in the
1285 vicinity of the southwestern block are composite and comprise 3500-3431 Ma granite,
1286 granodiorite, trondhjemite with minor tonalite, together with abundant 3350-3270 Ma tonalite
1287 to granodiorite, all of which in turn are intruded by 3230-3177 Ma diapiric trondhjemite plutons
1288 (this study; Jayananda et al., 2015; Ranjan et al., 2020). Inherited zircon xenocrysts from the
1289 present study indicate 3500 Ma, which is consistent with previous studies documented zircon
1290 ages of 3610 Ma (Guitreau et al., 2017). Granitoids adjoining to the southwestern block
1291 exhibits a wide range of compositions ($\text{SiO}_2 = 64.1 - 79.7$ wt.%, $\text{Al}_2\text{O}_3 = 10.7 - 17.3$ wt.% with
1292 a few mafic-rich layers displaying low SiO_2 in the range 52.0 - 60.6 wt.%). wt.%wt.%

1293 In an Ab-An-Or triangular diagram (O'Connor's, 1965; fields after Barker,
1294 1979), gneisses from the southwestern block plot on the trondhjemite, tonalite with a few
1295 samples extending into granite field (Fig.7.9a). These rocks can be further sub-divided into
1296 two major age groups: a younger 3350-3270 Ma group and an older 3500-3431 Ma group. The
1297 younger 3350-3271 Ma granitoids are dominantly tonalite to granodiorite with quartz, sodic
1298 plagioclase (An_{14-25}), minor K-feldspar, biotite, and occasional hornblende. They display

1299 variable Al contents (<14.55 to 17.30 wt.% Al₂O₃) and poorly to highly fractionated REE
1300 [(La/Yb)_N=8.51-33.74)] patterns (**Fig.7.10a**) with negative Ba, Nb, Ta, and Ti anomalies but
1301 variable Sr anomalies (**Fig.7.10b**) on multi-element diagrams. One biotite-rich mafic enclave
1302 (SiO₂ = 51 wt.%) with high total REE content (1154 ppm) display highly fractionated REE
1303 [(La/Yb)_N = 44.5) with negative Nb-Ta-Ba-Sr but positive Y anomaly on multi-element
1304 diagrams (figure not shown). In contrast the older (3500-3430 Ma) gneisses are dominantly
1305 trondhjemite to granite with quartz, sodic plagioclase (An₁₀₋₁₄), minor K-feldspar, biotite with
1306 occasional primary muscovite. They exhibit higher SiO₂ contents (>73 wt.%), lower Al₂O₃
1307 (<13.0 wt.%) but conversely higher Fe₂O₃ (0.4 - 4.8 wt.%) and MgO (0.43 to 1.43 wt.%). These
1308 older granitoids show low Ba-Sr, total REE (378-1211 ppm) with weakly to moderately
1309 fractionated REE patterns [(La/Yb)_N = 3.24-20.0)], with strong negative Eu anomalies (**see**
1310 **fig.7.10c**), but negative Ba, Nb, Ta, Sr, and Ti anomalies associated with positive Zr and Hf
1311 anomalies (**see fig.7.10d**).

1312 The TTG-type granitoids (3430 Ma) spatially associated with the northcentral block are
1313 tonalite-trondhjemite-granite in composition and contain quartz, sodic plagioclase (An₁₀₋₁₉),
1314 minor K-feldspar, biotite with occasional hornblende. They exhibit high SiO₂ (71-74 wt.%),
1315 low to high -Al (13.30-15.56 wt.%), low to moderate Ba, Sr, total REE (46-185 ppm), and
1316 poorly to moderately fractionated REE (**Fig.7.10e**) patterns [(La/Yb)_N = 1.06 -19.1]. They also
1317 show variable Eu anomalies (Eu/Eu* = 0.41 to 1.93) with negative Ba, Nb, Ta, Ti but negative
1318 to positive Sr anomalies on multi-element diagrams (**Fig.7.10f**).

1319 Granitoids from around the eastern block contain quartz, sodic plagioclase
1320 (An₁₅₋₂₃), minor K-feldspar, biotite and hornblende in mineral assemblages. Published data
1321 (**Jayananda et al., 2015; Ranjan et al., 2020**) for gneisses adjoining to the eastern block reveal
1322 younger ages of 3289-3276 Ma corresponding to a terminal event of TTG-type granitoids
1323 formation. The of studied samples have low-Al concentrations (<14.0 wt.%) except few

1324 samples show high-Al content (14.0 to 16.34wt.%) correlated with the SiO₂ content (64.14 to
1325 70.0 wt.%) with the exception of one sample that has 73.11 wt.% SiO₂. The low-Al samples
1326 have high-SiO₂ concentrations (73-76 wt.%), higher Fe₂O₃ and MgO, and they display low Ba-
1327 Sr, ΣREE (48-132 ppm), with poorly fractionated (**Fig.7.10g**) REE patterns [(La/Yb)_N = 9.62-
1328 14.09] and variable Eu (Eu/Eu* = 0.77-1.07) anomalies. These samples are also characterized
1329 by negative Ba, Nb, Ta, and Ti anomalies but positive Zr and Hf anomalies (**Fig.7.10h**). On
1330 the contrary, high-Al samples display high Ba-Sr, ΣREE (147-223 ppm) with highly
1331 fractionated REE patterns [(La/Yb)_N = 20.99-45.67] together with strongly negative Nb-Ta-Ti
1332 anomalies on multi-element diagram (**see fig.7.10g, h**).

1333 The 3230-3177 Ma diapiropic trondhjemites contain quartz, sodic plagioclase (An₁₀₋₁₂),
1334 minor K-feldspar, biotite and rare primary muscovite with accessory zircon, titanite and Fe-Ti
1335 oxides. They exhibit high-Al (>15.0 wt.%) except three samples which show low-Al (<14
1336 wt.%). Elemental characteristics like moderate to high Ba (371-724 ppm), Sr (322-735 ppm),
1337 low to moderate total REE (32-139 ppm) and moderate to highly fractionated REE [(La/Yb)_N
1338 = 5.27-20.03, **Fig.7.10i**), strong negative Nb-Ti anomalies but positive and negative Sr
1339 anomalies (**Fig.7.10j**).

1340 The other TTG-type granitoids found adjacent to the Nuggihalli, J.C.Pura,
1341 Ghattihosahalli, Banasandra, Nagamangala greenstone belts of the cratonic core show TTG
1342 affinity and contain quartz, sodic plagioclase (An₁₂₋₂₈), minor K-feldspar, biotite ± hornblende.
1343 They exhibit moderate to high SiO₂ (64.8-72.2 wt.%), high-Al (14.6-16.6 wt.%), low to
1344 moderate Ba, Sr, total REE (52-151 ppm), moderate to highly fractionated REE patterns
1345 (**Fig.7.10k**) with [(La/Yb)_N = 13.75-51.50] without any significant Eu anomalies except few
1346 samples show either negative or positive anomalies (Eu/Eu* = 0.72-1.22). They display
1347 negative Ba, Nb, Ta, Ti with flat or negative Sr anomalies on multi-element diagram

1348 (Fig.7.10). These granitoids exhibit ages in the range of 3400-3200 Ma (Jayananda et al., 2018,
1349 in prep; Ranjan et al., 2020; Bidyananda et al., 2016; Ao et al., 2021).

1350

1351 7.4 Zircon Lu-Hf isotopes

1352 Zircon populations for each sample show good consistency in their Lu-Hf isotope systematics
1353 (Table 1). Zircons from >3400 Ma samples have present-day $^{176}\text{Lu}/^{177}\text{Hf}$ that range from
1354 0.0003 to 0.0028 and present-day $^{176}\text{Hf}/^{177}\text{Hf}$ from 0.280612 to 0.280829, which give initial
1355 ϵ_{Hf} between +1.1 and +3.6. Zircons from 3350-3270 Ma samples exhibit present-day
1356 $^{176}\text{Lu}/^{177}\text{Hf}$ from 0.0008 to 0.0080 and present-day $^{176}\text{Hf}/^{177}\text{Hf}$ from 0.280769 to 0.281271,
1357 which give initial ϵ_{Hf} between +1.6 and +5.5 and three much higher datapoints with ϵ_{Hf} of +9.5,
1358 +11.1 and +11.7. The youngest group (3177 Ma) represented by one sample has present-day
1359 $^{176}\text{Lu}/^{177}\text{Hf}$ that range from 0.0011 to 0.0025 and present-day $^{176}\text{Hf}/^{177}\text{Hf}$ from 0.280833 to
1360 0.280935, which give initial $\epsilon_{\text{Hf(T)}}$ between +0.9 and +2.6. Overall, Lu-Hf isotope data show
1361 weakly to strongly depleted (compared to CHUR) signatures, hence, positive ϵ_{Hf} at the time of
1362 crystallization ($+1.9 \pm 0.6$ to $+4.9 \pm 0.6$; Fig.7.11a). Hafnium isotope signatures show a slight
1363 increase from 3430 to 3340 and then a decrease towards 3170 Ma, therefore, indicating tapping
1364 of an overall mildly depleted source, then a more depleted one that correspond to the backward
1365 projection of arc mantle present-day compositions (Dhuime et al., 2012; Iizuka et al., 2013),
1366 and finally a mildly depleted one again. Our new data are consistent with and complement
1367 literature data for igneous and detrital zircons, although the latter are not shown on Fig. 7.11a
1368 because detrital zircons commonly show much more variability than igneous crystals due to
1369 inherent impossibility to smooth out analytical scatter by considering multiple crystals or
1370 domains within crystals (e.g., Lancaster et al., 2015; Ranjan et al., 2022). As a consequence,
1371 we will only discuss Hf isotope data for igneous zircons to avoid artificial blurring of the
1372 magmatic signature of WDC granitoid formation.

1373

1374 **8. Petrogenesis of greenstone volcanic rocks**

1375 *8.1 Timing of greenstone formation*

1376 In the southwestern block, the contacts between greenstone ultramafic rocks and granitoids
1377 (TTG-type gneisses) are tectonized. It is difficult to provide precise ages for the ultramafic
1378 volcanism as $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of a few samples were affected by fluid flow associated with
1379 metamorphism and intrusion of ca. 3200 Ma diapiric trondhjemites. Ultramafic to mafic rocks
1380 of all of the three blocks together define an Sm-Nd whole rock isochron age of 3028 ± 53.3 Ma
1381 (MSWD = 2.3) which coincides with the cooling path of an amphibolite facies metamorphic
1382 event dated at 3091 ± 12 Ma (Jayananda et al., 2015). The ultramafic-mafic volcanic rocks of
1383 the southwestern block define a Sm-Nd whole-rock errorchron age of 3029 ± 91.5 Ma (MSWD
1384 1.2) which coincides with the cooling path of metamorphic event (Jayananda et al., 2013a).
1385 However, the occurrence of ultramafic-mafic volcanic rocks as large disrupted fragments
1386 within the 3400-3350 Ma granitoids of the southwestern block implies that volcanic rocks
1387 erupted prior to 3350 Ma.

1388 U-Pb zircon ages of this study together with published zircon ages (Jayananda et al.,
1389 2015; Guitreau et al., 2017; Ranjan et al., 2020) for adjacent granitoids reveal three major
1390 stages of formation. Zircon xenocrysts from trondhjemite to granite (Guitreau et al., 2017) and
1391 granodiorite from this study indicate a first stage of formation at 3600-3500 Ma. Our new U-
1392 Pb zircon data for tonalitic to granodioritic gneisses reveal a second episode at 3468-3410 Ma
1393 whilst zircon ages from several tonalitic to granodioritic samples indicate 3350-3300 Ma that
1394 correspond to a third stage of granitoid (TTG-type gneisses) formation.

1395 The ultramafic-mafic volcanic rocks of the northcentral block give an imprecise Sm-
1396 Nd regression age of 3124 ± 249 Ma (MSWD 2.1) which probably reflects partial resetting
1397 during metamorphic cooling. However, our U-Pb zircon data from a pyroclastic felsic flow

1398 from high stratigraphic levels with the Holenarsipur stratigraphy returned an age of 3305 ± 10
1399 Ma. The granitoids adjoining to the northcentral block yielded zircon U-Pb ages of 3429 ± 9
1400 Ma and 3429 ± 11 Ma without visible inherited cores. Stratigraphic relationships between
1401 northcentral block and adjoining granitoids remain uncertain due to tectonized contacts but
1402 the 3305 ± 10 Ma pyroclastic flow may represent a terminal stage of greenstone volcanism,
1403 hence, placing constraints on the timing of the northcentral block volcanism before 3305 Ma.

1404 The greenstone volcanic rocks of the eastern block provide a whole-rock Sm-Nd
1405 regression age of 3345 ± 111 Ma (MSWD=1.8) which is close to the U-Pb zircon age of 3347
1406 ± 11 Ma obtained for a plagiogranite that is interlayered with gabbroic rocks and that we
1407 interpret as a marker of a terminal stage of volcanism. Published U-Pb zircon ages for
1408 granitoids from the western margin of the Eastern block indicate younger ages of 3276 ± 5 Ma
1409 (Jayananda et al., 2015) and 3289 ± 5 Ma (Ranjan et al., 2020), consistent with a recent zircon
1410 age of 3242 ± 17 Ma (Panicker et al., 2021) as well as with a whole-rock Sm-Nd regression
1411 age of 3285 ± 170 Ma for an anorthosite body that intrudes the Eastern block (Bhaskar-Rao et
1412 al., 2000).

1413 Sm-Nd whole rock isotope data for ultramafic-mafic volcanic rocks from
1414 stratigraphically equivalent greenstone belts (i.e. Nuggihalli-Kalyadi, Banasandra and J.C.
1415 Pura) together with Holenarsipur volcanic rocks provide a whole-rock Sm-Nd regression age
1416 of 3356 ± 68 Ma which is assumed to represent an average age for widespread ultramafic
1417 volcanism (Jayananda et al., 2008), which is consistent with the timing of formation of volcanic
1418 rocks in the three blocks of the Holenarsipur greenstone belt.

1419

1420 8.2 Effects of alteration

1421 Elemental and isotopic data on komatiite and komatiitic-basalts from Archean cratons reveal
1422 mobility of some of the incompatible elements by fluid flow associated with metamorphism

1423 (Gruau et al., 1992; Lahaye et al., 1995; Polat et al., 2002; Polat and Hofmann, 2003;
1424 Chavagnac, 2004; Ordóñez-Calderón et al., 2008; Jayananda et al., 2008, 2016). The effect of
1425 low-grade metamorphism and hydrothermal alteration in element mobility in the Paleoproterozoic
1426 komatiites of the Western Dharwar craton have been discussed in detail elsewhere (e.g.
1427 Jayananda et al., 2008, 2016; Tushipokla and Jayananda, 2013). The studied komatiites and
1428 komatiitic-basalts of the Holenarsipur greenstone belt were affected by fluid flow associated
1429 with the intrusion of the 3200 Ma diapiric trondhjemite plutons as reflected in numerous
1430 quartzite veins with large tourmaline crystals traversing the volcanic rocks spatially associated
1431 with the plutons.

1432 Major elements (TiO_2 , Al_2O_3 , CaO , Fe_2O_3) together with trace elements Ni, Cr, V and
1433 Y define moderate to strong linear trends against MgO on the Harker binary diagrams which
1434 preclude extensive alteration of the studied greenstone volcanic rocks. However, large
1435 scattering of LIL elements (K, Rb, Ba, Sr) on the Harker binary plots (see fig.7.1d,e)
1436 nevertheless indicate their mobility during secondary processes, which is common in many
1437 komatiites (Arndt et al., 2008). The heat and fluids associated with the trondhjemite intrusions
1438 and following 3100 Ma metamorphic event (e.g., Rogers et al., 1986; Jayananda et al., 2013a,
1439 2015; Dasgupta et al., 2019) probably caused the mobility of large ion lithophile elements (K,
1440 Rb, Ba, Sr), Ce and Eu. On the other hand, the majority of samples show relatively uniform
1441 $\text{Al}_2\text{O}_3/\text{TiO}_2$ and $(\text{Gd}/\text{Yb})_N$ with smooth REE and primitive mantle normalized multi-element
1442 diagrams (see figs.7.6a).

1443 Incompatible element ratios such as Nb/U and Nb/Th are commonly used to evaluate
1444 the composition of source reservoirs of komatiites since these ratios are not significantly
1445 modified by magmatic processes (Campbell, 2002). Most of our data distribute continuously
1446 between MORB, primitive mantle (PM), and continental crust (CC) values (Sun and
1447 McDonough, 1989; Workman and Hart, 2005; Fig.8a) whilst some samples show excess

1448 scatter and even decoupling between Nb/U and Nb/Th. This is hard to reconcile with mantle
1449 processes and more likely reflect secondary alteration by fluids containing variable proportions
1450 of U and Th, these two elements being readily mobile in crustal fluids.

1451 Most $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ data for the studied volcanic rocks from three
1452 greenstone units define a linear array though with some scatter probably due to re-opening of
1453 Sm-Nd system coincide with LILE scattering. These elemental disturbances are probably a
1454 response to ca.3200 Ma trondhjemite intrusions and/or subsequent two metamorphic events at
1455 3100 Ma and 2500 Ma. To minimize the effects of hydrothermal alteration in the following
1456 discussion, we have used relatively immobile elements for which a magmatically-coherent
1457 behaviour was observed, as well as samples for which Sm-Nd isotope data appeared consistent
1458 with the main regression.

1459

1460 *8.3 Crustal contamination*

1461 Several attempts have been made to quantify the effect of crustal contamination in
1462 Archean komatiites and basalts (e.g. [Puchtel et al., 1997, 1998](#); [Polat et al., 1999, 2006](#)).
1463 Incompatible elements used to constrain source reservoirs of Archean komatiites are also
1464 sensitive to crustal contamination processes and to some extent affected by the degree of mantle
1465 melting (c.f [Puchtel et al., 1997, 1998](#); [Polat et al., 1999, 2006](#)). Therefore, it is important to
1466 characterize the role of crustal contamination while using incompatible elemental data in order
1467 to constrain mantle sources most rigorously. Incompatible element ratios are suitable indicators
1468 of crustal contamination as these elements are enriched in the continental crust. In particular,
1469 Zr/Th ratios are used to characterize crustal contamination of komatiites and basalts as
1470 continental crust is enriched in Th (Zr/Th = ~20) compared to the primitive mantle (Zr/Th
1471 =116; [Fan and Kerrich, 1997](#)). Field relationships in the Holenarsipur greenstone belt reveal

1472 that komatiites and komatiitic basalts erupted in marine environments as argued by visible
1473 pillow or flow-top pillow breccia structures. Most of the studied samples are not enriched in
1474 LIL elements and LREE but show flat REE patterns (see figs 7.2a-l). This together with high
1475 Zr/Th values (>80) preclude large-scale crustal contamination for the majority of studied
1476 komatiites and komatiitic basalts. A few samples from southwestern and northcentral blocks,
1477 nevertheless, show low Zr/Th values (60-70) which suggest minor ancient crustal
1478 contamination, or secondary enrichment of Th. Further, Kerrich et al. (1999) have used
1479 $(La/Sm)_N$ values to evaluate crustal contamination of Archean basalts from the Superior
1480 Province and showed that crustally contaminated basalts have high $(La/Ce)_N$ values (>1.5). In
1481 the studied greenstone volcanic rocks, a few samples ($\sim 10\%$) from the Holenarsipur,
1482 Nuggihalli, Kalyadi, J.C. Pura, and Nagamangala greenstone belts show $(La/Ce)_N$ values >1.5
1483 which indicate possible traces of crustal contamination. This is consistent with the observed
1484 Nb/Th values that also preclude extensive contamination by pre-existing crust (Polat et al.,
1485 2002; Polat and Hoffmann, 2003). Nd isotope data and Nb/Th ratios are not correlated (Fig.
1486 8b) which would be the case if Nd isotope signatures of the studied komatiites were controlled
1487 by crustal contamination with the least radiogenic being associated with the lowest Nb/Th
1488 ratios. This indicates that Nd isotope signature variability is either an artefact of secondary
1489 alteration or reflects source heterogeneity, or both. Most data presented in this contribution,
1490 either new or from the literature, can be used to discuss magmatic differentiation processes and
1491 composition of mantle source reservoirs.

1492 *8.4 Magmatic differentiation*

1493 The majority of greenstone volcanic rocks from the cratonic core are komatiite to komatiitic
1494 basalt in composition. Despite large variations in texture, mineralogy, major and trace element
1495 compositions, individual greenstones as well as greenstones altogether define moderate to
1496 strong linear trends on the Harker's binary diagrams (see fig.7.1d, e) which indicate that they

1497 share a common petrogenetic processes. Komatiites with more than 35 wt.% MgO possibly
1498 represent cumulate layers, whereas rocks with lower MgO content most likely represent liquid
1499 compositions. Among the greenstone volcanic rocks, Holenarsipur komatiites and komatiitic
1500 basalts from the three blocks plot co-linearly on binary plots of Al₂O₃, TiO₂, Fe₂O₃, CaO versus
1501 MgO suggesting their magmatic evolution by similar differentiation processes which
1502 essentially controlled by olivine fractionation (see fig.7.1d). Komatiites and komatiitic basalts
1503 from other greenstone belts in the cratonic core follow the same trends except few J.C. Pura
1504 komatiitic basalt samples which seem to have additional minerals (i.e., olivine, pyroxene,
1505 feldspar, ilmenite) controlling their compositions as illustrated in MgO versus SiO₂ or MgO
1506 versus TiO₂ plots (see fig.7.1d). Among trace elements, Ni and Cr exhibit positive correlations
1507 with MgO which suggest olivine and possible chromite fractionation, also consistent with the
1508 negative correlations between Zr, Y, V and MgO which likely reflect enrichment by fractional
1509 crystallization or increasing depletion concomitant with higher degrees of partial melting.
1510 Commonly flat REE and trace element patterns indicate large degrees of partial melting
1511 (>40%). Negative Ti anomalies on the primitive mantle normalized multi-element patterns
1512 coupled with negative correlation of TiO₂ with MgO on the Harker binary plots very likely
1513 reflect a control by accessory phase rutile.

1514 *8.5 Source characteristics and petrogenetic processes*

1515 Komatiites and komatiitic basalts from the Archean greenstone belts are key to our
1516 understanding of thermal and chemical evolution of the mantle, continental growth and
1517 geodynamics of the early earth (Polat et al., 1999, 2006; Arndt et al., 2008; Herzberg et al.,
1518 2010; Mole et al., 2015; McKenzie, 2020). Despite numerous studies on the Archean komatiites
1519 involving petrologic, elemental, isotopic, experimental and simulation models, the origin of
1520 komatiite magma is still subject of much discussion on whether the magma originated by
1521 anhydrous melting of deep mantle (for reviews see Arndt et al., 2008; Herzberg, 1995;

1522 [Herzberg et al., 2010](#); [Mc Kenzie, 2020](#)) or wet melting of the shallow mantle ([Parman et al.,](#)
1523 [1997](#)), or deep hydrous mantle reservoirs ([Sobolev et al., 2019](#)). Archean komatiite flows are
1524 characterized by high MgO content (18-35 wt.%) excluding cumulate layers which contain
1525 >35.0 wt.%, but conversely low Al₂O₃ concentrations and very low incompatible element
1526 contents. These characteristics are attributed to high-degree (40-50%) partial melting of deep
1527 mantle reservoirs ([Arndt, 2003](#); [Sossi et al., 2016](#)). Archean komatiites are commonly divided
1528 into two principal types including the Al-depleted and Al-undepleted ([Arndt and Nisbet, 1982](#)).
1529 Subsequently, [Arndt \(2003\)](#) proposed Al-depleted komatiites to ca.3500 Ma Barberton-type
1530 with sub-chondritic Al₂O₃/TiO₂ (<16), CaO/Al₂O₃ (>1.0), fractionated REE patterns with high
1531 (Gd/Yb)_N (>1.0), whilst Al-undepleted komatiites to 2700 Ma Munro-type (Abitibi-type)
1532 characterized by near chondritic Al₂O₃/TiO₂ (≤20), CaO/Al₂O₃ (<1.0) and poorly fractionated
1533 with (Gd/Yb)_N (<1.0). Incompatible trace element ratios like Zr/Nb, Nb/Th, Nb/Yb, Th/Yb,
1534 Zr/Y, Nb/Y, (Gd/Yb)_N, La/Yb and Hf/Sm are extensively used to evaluate their mantle source
1535 reservoirs, mineralogy of melt residue, depths and tectonic contexts of melt generation (e.g.
1536 [Fitton et al., 1997](#); [Baksi, 2001](#); [Condie, 2003](#); [Chavagnac, 2004](#); [Sossi et al., 2016](#); [Jayananda](#)
1537 [et al., 2016](#); [Wyman, 2020](#)). Low Zr/Nb and Hf/Sm together with high Nb/Y and La/Yb in the
1538 Al-depleted Archean komatiites are usually attributed to the existence of depleted reservoirs
1539 and recycled materials along with enriched components in the deep mantle by 3500 Ma
1540 ([Kellogg et al., 1999](#); [Kerrick and Xie, 2002](#); [Polat and Kerrich, 2000](#)). Further, low Zr/Nb and
1541 Hf/Sm ratios coupled with high La/Yb and Nb/Y values in the Al-depleted komatiites may be
1542 linked to majorite fractionation, whilst high Zr/Nb and Nb/Y ratios in Al-undepleted komatiites
1543 could be related to melting of shallow mantle source with garnet entering into melt. [Sossi et al](#)
1544 [\(2016\)](#) while addressing the petrogenesis of komatiites from the five Archean cratons argued
1545 that Al-depleted Barberton type komatiites were generated by high-degree batch melting (~
1546 40%,) at high pressure (9.0 GPa) and high temperature (T_p = 1950°C), hence, in the deep

1547 mantle where majorite can be stable in the melting residue. In contrast, Al-undepleted Munro-
1548 type komatiites originated in a shallower mantle by moderate pressure fractional melting (~
1549 25%, 5.0 GPa, $T_p = 1750^\circ\text{C}$) as density contrast between the liquid and solid increase at such
1550 depths. More recent thermodynamic model has shown that komatiite melts originate at great
1551 depth (~600 km) in a rising plume where Ca-perovskite is stable as a residual phase, therefore,
1552 accommodating a fair amount of trace elements which, in turn, accounts for low contents of
1553 incompatible elements in generated komatiitic magmas (McKenzie, 2020).

1554 The 3400-3200 Ma old komatiite and komatiitic basalts from the cratonic core in the
1555 WDC show significant variation in elemental compositions and Nd isotope signatures
1556 suggesting distinct mantle source reservoirs, depth and degree of melting and tectonic context
1557 of melt generation.

1558 *8.5.1 The Holenarsipur greenstone volcanic rocks – the three blocks*

1559 The petrogenetic interpretations were made on the samples devoid of evident alteration and
1560 weathering. However, exact petrological type of ultramafic-mafic rocks from three blocks may
1561 not be good indicators of their origin, but geochemical signatures can help constrain source
1562 reservoirs and unravel the geodynamic context of formation of such rocks (e.g., Xie and
1563 Kerrich, 1994, 1995; Puchtel et al., 1997, 1998; Polat et al., 1999, 2006; Pearce, 2008; Sobolev
1564 et al., 2016). Major elements of three greenstone units plotted on triangular diagrams (Al_2O_3 -
1565 $\text{Fe}_2\text{O}_3+\text{TiO}_2$ -MgO; Jensen, 1976) and (CaO -MgO- Al_2O_3 ; Viljoen et al., 1982) reveal dominant
1566 komatiite to komatiitic basalt composition although with some distinctions in trace element
1567 signatures indicating independent histories.

1568 The southwestern block has an affinity with an oceanic plateau with remnants of
1569 microcontinent. Incompatible element contents (REE, Nb, Y, Th, Zr; Figs. 7.2a-d) reveal two
1570 groups. The low REE group ($\Sigma\text{REE} = 4.4$ -18.8 ppm with lower SiO_2 44.1-46.0 wt.%) exhibit

1571 flat to slightly fractionated REE patterns (**Fig. 7.2a**) with $(\text{Gd}/\text{Yb})_{\text{N}} = 0.53\text{-}1.86$ which suggest
1572 their derivation by high-degree partial melting of a shallow depleted mantle, hence, without
1573 residual garnet. This is consistent with the absence of Nb anomalies together with positive Zr
1574 and Y anomalies on primitive mantle normalized multi-element diagram (**Fig. 7.2b**). The high-
1575 REE group ($\Sigma\text{REE} = 20.6\text{-}66.7$ ppm) with fractionated REE [$(\text{Gd}/\text{Yb})_{\text{N}} 1.22\text{-}2.29$ (**Fig. 7.2c**)
1576 combined with positive Nb, coupled negative Zr and Hf anomalies on multi-element diagrams
1577 (**Fig. 7.2d**) suggest origin of melts from deeper primitive mantle reservoirs with possible
1578 residual garnet (majorite). Sm-Nd isotope data of the komatiites were used to calculate $\epsilon_{\text{Nd}(\text{T})}$
1579 at 3350 Ma to constrain mantle sources and possible crustal contamination. Nine samples with
1580 large range of values ($\epsilon_{\text{Nd}(\text{T})} = -0.1$ to $+4.6$) suggest chondritic to depleted (MORB-like) mantle
1581 reservoirs whilst two samples with negative ($\epsilon_{\text{Nd}(\text{T})} = -0.8$ to -1.6) values are likely related to
1582 possible contamination of pre-existing crust or artefact of secondary modification of the Sm-
1583 Nd system (Meen et al., 1992). Plots of incompatible element ratios like Th/Yb versus Nb/Yb
1584 (**Fig. 8c**; Van Kranendonk et al., 2015) and Nb/Y versus Zr/Y (**Fig. 8d**; Condie, 2003) suggest
1585 primitive to depleted mantle sources whilst Zr/Nb versus Nb/Th plot (**Fig. 8e**; Condie, 2003)
1586 attribute their eruption as oceanic plateaus. The above lines of arguments for oceanic plateau
1587 origin of southwestern block komatiites are consistent with studies on komatiites from other
1588 cratons (e.g., Desrochers et al., 1993; Storey et al., 1991; Kusky and Kidd, 1992).

1589 The northern block is likely an oceanic arc with coeval mini volcanic plateau fragments.
1590 Major element data ($\text{SiO}_2 = 45.1$ to 68.6 wt.%; $\text{MgO} = 23.9$ to 9.7 wt.%) reveal komatiite,
1591 komatiitic basalt, basalt to dacite compositions. Komatiites contain low to moderate ΣREE
1592 ($14.6\text{-}59.1$ ppm), display flat to fractionated REE patterns with $(\text{Gd}/\text{Yb})_{\text{N}}$ ratios of $0.99\text{-}1.68$
1593 (**Fig. 7.2e**) whilst komatiitic basalts show moderate to high ΣREE ($27.3\text{-}138.0$ ppm) and display
1594 fractionated REE patterns [$(\text{Gd}/\text{Yb})_{\text{N}} = 0.94\text{-}1.93$] (**Fig. 7.2g**), hence, implying heterogeneous
1595 mantle reservoirs whilst negative Nb anomalies coupled with either positive or no significant

1596 Zr, Hf and Y anomalies on the multi-element diagrams (**Fig.7.2f and h**) suggest an origin from
1597 shallower depths. These elemental characteristics, lithological association of komatiite-
1598 komatiitic basalt- basalt- dacite points to a possible oceanic arc environment. Incompatible
1599 element ratios like Th/Yb versus Nb/Yb and Nb/Y versus Zr/Y (**see fig.8c, d**) indicate
1600 heterogenous sources involving primitive to enriched MORB mantle, with few samples
1601 showing modern N-MORB signatures, whilst on Zr/Nb versus Nb/Th studied samples
1602 essentially fall within the field of arc magmas. These geochemical characteristics point to a
1603 possible oceanic arc origin, nevertheless with local mantle heterogeneities (**see fig. 8e**). Initial
1604 Nd isotope ratios for the ten samples gave with ($\epsilon_{Nd(T)=3350 \text{ Ma}}$) values ranging from -0.4 to
1605 +4.5, except one sample having -2.6, which suggest chondritic to depleted mantle sources with
1606 possible traces of ancient crustal contamination and/or secondary alteration for the negative
1607 $\epsilon_{Nd(T)}$ value.

1608 The eastern block resembles an oceanic crustal section close to a spreading centre. The
1609 komatiitic to basaltic volcanic rocks with SiO₂ ranging from 42.7 to 50.0 wt.% form two
1610 groups: low Σ REE and high Σ REE. The low Σ REE (5.80-22.20 ppm) group displays flat
1611 patterns with depletion in LREE and poorly fractionated [$(Gd/Yb)_N = 0.61-1.38$] suggest a
1612 depleted mantle reservoir similar to N-MORB source (e.g., Xie and Kerrich, 1994, 1995;
1613 Puchtel et al., 1997, 1998; Polat et al., 1999, 2006). The absence of Nb anomalies, coupled with
1614 positive Zr, Y anomalies on the multi-element diagram (**Fig. 7.2i, j**) point to melt origin from
1615 a shallow depleted mantle which is consistent with $\epsilon_{Nd(T)}$ of +2.46 to +5.5 at 3350 Ma. On the
1616 contrary, the high Σ REE (48.39-294.79 ppm) group shows fractionated REE patterns
1617 [$(Gd/Yb)_N = 1.16-1.96$] coupled with positive Nb anomalies and negative Hf-Y anomalies on
1618 the multi-element diagrams (**Fig. 7.2k, l**) which coupled with $\epsilon_{Nd(T)}$ of -0.8 to +0.1 at 3350 Ma
1619 suggest a primitive mantle source. Elemental ratios like Th/Yb versus Nb/Yb, and Nb/Y versus
1620 Zr/Y, indicate their origin from essentially N-MORB to E-MORB sources (**Fig.8c, d**).

1621 Moreover, on Zr/Nb versus Nb/Th plot suggest that volcanic rocks of the eastern block
1622 originated from different depths in the mantle with affinities to modern N-MORB to E-MORB,
1623 hence, pointing to a likely preserved oceanic crust (Xie and Kerrich, 1995) close to a spreading
1624 centre (**Fig. 8e**).

1625

1626 *8.5.2 Stratigraphically equivalent greenstone belts in the cratonic core*

1627 The Nuggihalli-Kalyadi volcanic assemblages are mainly komatiitic sharing similar
1628 compositional characteristics (low SiO₂ = 38.9-46.9 wt.%; high MgO = 30.9-34.9 wt.%;
1629 CaO/Al₂O₃ = 0.63-1.32; Al₂O₃ /TiO₂ = 9.78-23.48), high compatible element contents (Ni =
1630 1388-1960 ppm; Cr = 2209-3859 ppm) and variable depletion in incompatible trace elements
1631 (Σ REE = 4.2-30.9 ppm). Their trace element ratios (e.g., (Gd/Yb)_N = 0.43-1.24) coupled with
1632 sub-chondritic to moderate total REE contents, suggest their origin from heterogeneous mantle
1633 (highly depleted to primitive mantle) reservoirs at different depths. The flat REE patterns
1634 coupled with positive Zr, Ti and Y anomalies (two samples with negative anomalies) on PM
1635 normalized multi-element diagrams are consistent with their origin by high-degree melting (40-
1636 50%) of mantle at different depths, hence, with possible influence of residual garnet.
1637 Incompatible element ratios like Th/Yb versus Nb/Yb (**see fig.8c**) coupled with Sm/Nd versus
1638 Nb/Th reveal highly depleted to primitive mantle sources which is consistent with $\epsilon_{Nd(T)}$ of
1639 +0.5 to +5.4. These observations altogether suggest dominant depleted mantle reservoirs for
1640 the source of the Nuggihalli-Kalyadi volcanic rocks.

1641 The J.C. Pura greenstone volcanic assemblages are dominantly komatiite to komatiitic basalt
1642 flows with massive cumulate layers. These thick lava flows along with cumulates were
1643 interpreted as differentiation sections of a volcanic plateau (Jayananda et al., 2016). The
1644 komatiite flows contain high MgO contents (22.3-31.9 wt.%), low SiO₂ (45.1-47.2 wt.%),

1645 variable $\text{CaO}/\text{Al}_2\text{O}_3$ (0.70-1.66) and $\text{Al}_2\text{O}_3/\text{TiO}_2$ (5.38-22.47), low to moderate total REE
1646 (14.06-45.92 ppm), whilst komatiitic basalts show lower MgO (12.9-15.3 wt.%), higher SiO_2
1647 (49.0-50.5 wt.%), variable $\text{CaO}/\text{Al}_2\text{O}_3$ (0.77-1.30), and slightly higher $\text{Al}_2\text{O}_3/\text{TiO}_2$ (15.5-20.8).
1648 In contrast, the cumulate layers are characterized by higher MgO contents (35.1-42.2 wt.%)
1649 and low SiO_2 (41.1-45.4 wt.%), $\text{CaO}/\text{Al}_2\text{O}_3$ (0.70-1.66), and $\text{Al}_2\text{O}_3/\text{TiO}_2$ (5.38-22.47). The
1650 moderate to strong linear trends between TiO_2 , Al_2O_3 , Fe_2O_3 and CaO in Harker's binary
1651 diagrams indicate fractionation of olivine, \pm pyroxene, \pm ilmenite (see fig.7.1d). Plots of
1652 $\text{CaO}/\text{Al}_2\text{O}_3$ versus $(\text{Gd}/\text{Yb})_N$ (see fig.7.6a) and $\text{Al}_2\text{O}_3/\text{TiO}_2$ versus $(\text{Gd}/\text{Yb})_N$ (see fig.7.6b)
1653 suggest that komatiitic magmas originated from various mantle reservoirs at different depths
1654 similar to the Barberton and Abitibi-type komatiites, with most data belonging to the former
1655 type. It is, however, hard to reliably discriminate source characteristics from effects of
1656 secondary alteration and/or fractional crystallization on these elemental ratios. The sub-
1657 chondritic to moderate total REE (3.6-57.4 ppm) coupled with $\epsilon_{\text{Nd}(T)}$ values between +0.50 and
1658 +5.4 at 3350 Ma reveal their origin from primitive to depleted (N-MORB like) mantle
1659 reservoirs. This is also reflected on incompatible trace element ratios projected on the Th/Yb
1660 versus Nb/Yb (see fig.8c) and the Nb/Y versus Zr/Y plots (see fig.8d). The lithological
1661 associations (voluminous and thick komatiite and komatiitic basalt flows forming pillows with
1662 flow top breccias coupled with cumulates) suggest their eruption in an oceanic plateau
1663 environment. This conclusion is consistent relationships within Zr/Nb versus Nb/Th plot
1664 (Condie, 2003) suggesting eruption in oceanic plateau setting (see fig.8e).

1665 Among Ghattihosahalli volcanic rocks, voluminous komatiite flows with cumulate layers
1666 (Jayananda et al., 2008) together with associated fuchsite quartzite, barite beds imply an
1667 oceanic setting. Abundant serpentine, highest MgO (39.3-41.3 wt.%) and Ni (3145-3177 ppm)
1668 coupled with low total REE (9.76-9.81 ppm) suggest a cumulative nature. Their low
1669 incompatible element contents along with their incompatible trace element ratio plots (Th/Yb

1670 vs Nb/Yb and Nb/Y vs Zr/Y), $\epsilon_{\text{Nd(T)}} = +2.8$ to $+5.3$ (see fig.8c, d) reveal significantly depleted
1671 mantle sources. Their high MgO contents, fractionated REE patterns [$(\text{Gd/Yb})_{\text{N}} = 2.26-2.52$]
1672 together with their high $\text{CaO}/\text{Al}_2\text{O}_3$ (1.32-1.34) and low $\text{Al}_2\text{O}_3/\text{TiO}_2$ (average of 5) argue for
1673 an origin of komatiite magmas by high-degree melting ($>40\%$) in the deep mantle (>400 km)
1674 with garnet stable in the melting residue. Elemental ratios like Zr/Nb and Nb/Th (see fig.8e)
1675 together with lithological associations (komatiite-fuchsite quartzite-barite) point to an eruption
1676 of deep-seated melts in an oceanic plateau setting.

1677 Among the Banasandra volcanic rocks, voluminous komatiites displaying cumulate, spinifex,
1678 pillow to pillow breccias with hyaloclastite structures suggest their eruption in shallow marine
1679 environment to slightly above sea level. Their elemental characteristics such as high MgO
1680 (30.3-42.4 wt.%), sub-chondritic to low total REE (3.5-12.2 ppm), Nd isotopes ($\epsilon_{\text{Nd(T)}} = +0.1$
1681 to $+5.1$ excluding two anomalous values), coupled with flat REE patterns suggest high-degree
1682 partial melting of a heterogeneous mantle (primitive to highly depleted) reservoirs which is
1683 also reflected in plots of incompatible element ratios like Th/Yb versus Nb/Yb (see fig.8c) and
1684 Nb/Y versus Zr/Y plots (see fig.8d). The observed $\text{CaO}/\text{Al}_2\text{O}_3$ (0.6-1.2), $\text{Al}_2\text{O}_3/\text{TiO}_2$ (8.2-18.5)
1685 and $(\text{Ga/Yb})_{\text{N}} = 0.99-1.90$ are compatible with an origin of komatiitic magmas from a deep
1686 mantle reservoirs retaining variable garnet (majorite?) in the melt residue. Field evidences of
1687 thick peridotitic komatiite flows forming pillow and flow-top breccias, spinifex with cumulate
1688 layers combined with incompatible element ratios such as Zr/Nb versus Nb/Th (see fig.8e)
1689 suggest an oceanic plateau setting.

1690 The Nagamangala komatiites exhibit crude pillow and spinifex structures as well as low SiO_2
1691 contents (40.5-43.9 wt.%), conversely high MgO (28.8-35.8 wt.%), Ni (1220-1589 ppm), Cr
1692 (1294-3565 ppm) but lower incompatible elements ($\Sigma\text{REE} = 9.2-20.9$ ppm; Th = 0.14-0.54
1693 ppm; Nb = 1.1-1.25 ppm; Zr = 6.9-20.09 ppm; Hf = 0.04-0.05 ppm; Tushipokla and Jayananda,

1694 2013). The majority of studied samples belong to Al-depleted Barberton type komatiites
1695 ($\text{CaO}/\text{Al}_2\text{O}_3 = 1.02 - 1.18$; $\text{Al}_2\text{O}_3/\text{TiO}_2 = 11.69 - 16.86$; $(\text{Gd}/\text{Yb})_N = 0.59 - 1.46$) whilst other
1696 samples have Al-undepleted Munro-type affinity ($\text{CaO}/\text{Al}_2\text{O}_3 = 0.67 - 0.98$ and $\text{Al}_2\text{O}_3/\text{TiO}_2 =$
1697 $11.37 - 19.70$; $(\text{Gd}/\text{Yb})_N < 0.9$). The Al-undepleted character ($\text{CaO}/\text{Al}_2\text{O}_3 < 0.8$), despite high
1698 MgO, low $\text{Al}_2\text{O}_3/\text{TiO}_2 (< 15)$ and high $(\text{Gd}/\text{Yb})_N > 1.0$, attributed to a possible loss of CaO in
1699 komatiites by post-magmatic fluid driven hydrothermal processes (Chikhaoui., 1981;
1700 Tushipokla and Jayananda, 2013). The low to moderate incompatible element contents (ΣREE
1701 $= 9.24 - 50.98$) in both Al-depleted and Al-undepleted komatiites together with their flat REE
1702 patterns and high MgO can be attributed to high-degree (40-50%) melting of deep mantle
1703 sources. The slight positive or negative Nb anomalies on multi-element diagrams are attributed
1704 plume source with possibility of recycled slab component involvement (Tushipokla and
1705 Jayananda, 2013). Incompatible element plots of Th/Yb vs Nb/Yb, Nb/Y vs Zr/Y and Nb/Th
1706 versus Nb/U also suggest depleted to primitive mantle sources, with a possible volcanic plateau
1707 environment (see figs.8a-d).

1708 To summarize typical geochemical fingerprints of subduction zone such as enrichment
1709 of LIL elements, depletion of HFSE (Nb, Ta, Zr, Hf, Ti, Y and Th) compared to REE are not
1710 observed in most of komatiites and komatiitic basalts except in komatiitic basalts of the
1711 Northcentral block of the Holenarsipur schist belt. This precludes the involvement of shallow
1712 arc mantle and argue for deep mantle sources for most komatiites of the cratonic core.

1713

1714 **9. Origin of TTG-type granitoid and diapiric trondhjemites**

1715 9.1 *Around the southwestern block (3430-3270 Ma granitoids with 3500-3600 Ma remnants)*

1716

1717 Major element compositions of the granitoids when projected on normative Ab-An-Or
1718 triangular plot (O'Connor, 1965) reveals dominant trondhjemite to tonalite composition with

1719 few samples extending granodiorite and granite field, (see fig.7.9a). On Q-Ab-Or triangular
1720 diagram (Barker and Arth, 1976) these gneisses deviate from classical calc-alkaline
1721 differentiation trend and distribute around the trondhjemite differentiation trend (Fig.7.9b)
1722 typical of Archean TTG suites (Martin, 1994). Considering age and geochemical
1723 characteristics, these granitoids can be grouped into two major groups, as presented in the result
1724 section, a high-Al and a low-Al group. Major elements show consistent trends between the two
1725 groups which suggest possible control by hornblende, plagioclase, garnet, titanite, ilmenite and
1726 apatite (Fig. 7.9c). Trace element patterns further confirm the control by rutile in both groups
1727 as argued by the negative anomalies of Nb, Ta, and Ti (Fig. 7.10). The influence of feldspar is
1728 confirmed in the low-Al group that exhibit strongly negative, Ba, Sr, and Eu anomalies but it
1729 is less clear in the high-Al group that only show very weak Eu, Ba, and Sr anomalies that are
1730 both positive and negative (Fig. 7.10). This latter group also exhibits a small, but visible
1731 positive Zr-Hf anomaly that may be due to zircon fractionation. Since low-Al gneisses are also
1732 characterized by high SiO₂ concentrations, it is hard to discriminate the influence of the source
1733 from that of fractional crystallization on rock chemical compositions. Therefore, the chemistry
1734 of these gneisses is likely dominated by fractional crystallization of the above-mentioned
1735 mineral phases, superimposed on source characteristics. Geochemical data are, here,
1736 interpreted as representative of magmatic liquids, as commonly done in the literature, but it
1737 should be kept in mind that the granitoids rarely represent pure magmatic liquids and
1738 cumulative effects can also add to factors influencing rock compositions. The high-Al group
1739 exhibits a larger range of SiO₂ than the low-Al group and it is, hence, easier to discriminate
1740 between source and fractional crystallization effects. For instance, the anomalies visible for
1741 Ba, Sr, Eu, Zr, and Hf are weak and both positive and negative (Fig. 7.10) which most likely
1742 indicate that they were developed in some samples in response to mineral fractionation during
1743 crystallization. On the contrary, Nb, Ta, and Ti anomalies are shared by all samples regardless

1744 of their SiO₂ content which suggest a source control, more or less accentuated by fractional
1745 crystallization.

1746 The general shape of REE (La/Yb)_N and trace element patterns indicate that the high-
1747 Al group is more fractionated than the low-Al one which is most likely a primary feature of
1748 these rocks. This is because fractional crystallization of granitoid magmas corresponding to
1749 grey gneiss compositions commonly involve plagioclase and hornblende fractionation (e.g.,
1750 [Martin, 1986](#); [Laurent et al., 2020](#); [Kendrick et al., 2021](#)), the latter of which imposes those
1751 fractionated liquids should have lower HREE contents and, hence, more fractionated REE
1752 patterns. Therefore, the higher HREE contents and less fractionated REE patterns of the low-
1753 Al group seem to be a primary feature of this group which, consequently, indicate that the high-
1754 and low-Al gneisses cannot be genetically related. Elemental ratio plots (i.e., Nb/Ta versus
1755 Gd/Yb; Nb/Ta versus Zr/Sm; Sr/Y versus K₂O/(Na₂O+CaO)) indicate the involvement of
1756 mineralogical phases consistent with those deduced from separate major and trace element data
1757 and allow further discussion about the sources of these granitoids ([Fig. 9a-c](#)). Data plotted in
1758 [figures 9a-9f](#) show significant scatter that can be interpreted as resulting from fractional
1759 crystallization, different depths and mineralogy of melting residues or both. It is difficult to
1760 discriminate the influence of each parameter separately but we can conservatively conclude
1761 that the source lithologies of granitoid gneisses around the southwestern block were
1762 geochemically enriched mafic rocks ranging from low-K to high-K basaltic to tonalitic in
1763 composition and that they melted at different depths from low pressures for the low-Al group
1764 to high pressures for the high-Al group, as notably, but not only, argued by La/Yb and Sr/Y
1765 ratios ([see figs. 7.10 and 9](#)).

1766 Whole-rock Nd isotope signatures together with zircon Hf isotope compositions ([Fig](#)
1767 [7.11a, b](#)) indicate dominant mildly to significantly depleted sources for the granitoids around
1768 the southwestern block. Hafnium isotopes exhibit less scatter than Nd isotopes likely because

1769 the Sm-Nd isotope system can be more readily disturbed at the whole-rock scale than the Lu-
1770 Hf isotope system within zircon. Consequently, most source compositions are discussed based
1771 on the Hf isotope data. Since the granitoids cannot be directly extracted from the mantle for
1772 geochemical mass balance reasons, Hf isotopes in zircon trace the isotope composition of a
1773 mafic precursor, which was identified based on elemental geochemistry (**Fig. 9**), reveals that
1774 the source was previously extracted from the mantle. Data of granitoid gneisses sampled
1775 around the southwestern block from this study and from the literature are consistent with each
1776 other and show mantle-like signatures (**Fig. 7.11a**) which indicate that the crustal residence
1777 time of these mafic precursors were short, which was also concluded by [Guitreau et al., 2017](#)
1778 and [Ranjan et al., 2020](#) for WDC Peninsular gneisses. A notable feature of Hf isotope data is
1779 the slightly less depleted nature of >3400 Ma granitoids compared to the 3350-3270 Ma group
1780 that, in contrast, exhibit consistently identical signature on this time range. Also note the fairly
1781 good agreement between backward projected MORB and arc mantle signatures and those of
1782 studied granitoid gneisses (**Fig. 7.11a**).

1783

1784 *9.2 Around the northcentral block (3430-3302 Ma granitoids)*

1785 Granitoids adjoining to the northcentral block are tonalite-trondhjemite-granite in composition
1786 and contain quartz, sodic plagioclase (An₁₀₋₁₉), minor K-feldspar, biotite with occasional
1787 hornblende. Geochemically, these granitoids are very similar to the high-Al group from the
1788 southwestern block (e.g., high (La/Yb)_N, negative Nb-Ta-Ti anomalies, and variable Ba, Sr,
1789 and Eu anomalies) except that they have slightly more variable total REE concentrations, as
1790 well as Sr, Eu and Ta anomalies, most likely resulting from more visible effects of fractional
1791 crystallization (**Figs. 7.10 and 9**). Moreover, Hf and Nd isotopes are also similar between
1792 granitoids found around the northcentral and the southwestern block (**Fig. 7.11**). As a

1793 consequence, they can be interpreted in the same way as the high-Al group discussed in the
1794 previous section and are perhaps local equivalents.

1795

1796 *9.3 Around the eastern block (3289-3276 Ma granitoids)*

1797 These granitoids contain quartz, sodic plagioclase (An₁₅₋₂₃), minor K-feldspar, biotite and
1798 hornblende in mineral assemblage. Published data (Jayananda et al., 2015; Ranjan et al., 2020;
1799 Ao et al., 2021) for gneisses adjoining to the eastern block reveals younger ages of 3289 -3276
1800 Ma that correspond to a terminal event of granitoid formation around the Holenarsipur
1801 greenstone belt. They have high-Al concentrations (14.98-16.34 wt.%) with significant
1802 variation in SiO₂ (64.14-73.11 wt.%). These granitoids have geochemical characteristics
1803 identical to the high-Al group found around the southwestern block (see section 9.2)
1804 Furthermore, some granitoids around the eastern block show more pronounced positive Zr-Hf
1805 anomalies coupled with positive Y anomalies which are, altogether, most likely related to
1806 zircon accumulation in these rocks. Isotopic data also indicate similar sources for all the
1807 granitoid gneisses regardless of their geographic location, within the limits of uncertainty
1808 considered. (Fig. 7.11a-b).

1809

1810 *9.4 Diapiric trondhjemites (3230-3180 Ma plutons)*

1811 Our new zircon U-Pb ages combined with published data indicate 3230-3177 Ma for intrusion
1812 their magmatic precursors (Jayananda et al., 2015; Guitreau et al., 2017; Ranjan et al., 2020).
1813 The mineral assemblages comprise quartz, sodic plagioclase (An₁₀₋₁₂), minor K-feldspar,
1814 biotite and primary muscovite with accessory zircon, titanite and Fe-Ti oxides. They exhibit
1815 high-Al contents (>15.0 wt.%) except for three samples that show low-Al (<14 wt.%). These
1816 granitoids share many characteristics similar with the high-Al group gneisses around the
1817 southwestern and northcentral blocks (e.g., Al₂O₃ and SiO₂ concentrations, high (La/Yb)_N,

1818 variable Ba, Eu, and Sr anomalies, and negative Ti anomalies) but differs from these latter by
1819 the absence of Nb-Ta anomalies and, on the contrary, positive U and Th anomalies (**Fig. 7.10**).

1820 These rocks seem to differ from other granitoids (TTG-type gneisses) in terms of their sources
1821 as the $3 \cdot \text{CaO} - \text{Al}_2\text{O}_3 / \text{FeO}_t + \text{MgO} - 5 \cdot \text{K}_2\text{O} / \text{Na}_2\text{O}$ triangular plot ([Laurent et al., 2014](#)) indicate a
1822 dominant tonalitic source (**see fig. 9d**). However, if the whole rock chemistry particularly
1823 elemental ratios (Nb/Ta, Zr/Sm, Sr/Y, La/Yb; see fig. 9a, b, c and e) is interpreted in terms of
1824 melting pressures, the trondhjemites require moderate to high pressures similar to the high-Al
1825 group. Therefore, the magmatic precursors of these granitoids could be more evolved than that
1826 of other granitoids, hence, accounting for their slight geochemical differences.

1827 Neodymium and Hf isotopes of the trondhjemites indicate a less depleted source compared to
1828 3350-3270 Ma granitoids but more similar to the >3400 Ma group (**Fig. 7.11a-b**). This could
1829 be due to longer crustal residence times of their magmatic precursors compared to those of the
1830 other granitoid groups or, alternatively, their magmatic precursors being more evolved, their
1831 Lu/Hf would be lower than that of the other granitoids precursors, hence, resulting in a more
1832 rapid drift from mantle signatures (**see fig 7.11a-b**). These trondhjemite plutons could actually
1833 be the reworking products of tonalites equivalent to 3350-3270 Ma granitoids, which would be
1834 realistic both in terms of elemental and isotopic geochemistry (**Figs. 7.10 and 7.11a-b**).

1835 Alternatively, their elemental and isotopic signatures could be explained by mixed sources
1836 involving mafic crustal lithologies, akin to those of the other granitoid groups, contaminated
1837 by felsic crustal lithologies (i.e., >3270 Ma gneisses). We cannot objectively exclude either of
1838 these hypotheses, but this does not affect the geodynamic model proposed in this study.

1839 Trondhjemite plutonism was spatially linked with a major phase of crustal-scale partial
1840 convective overturn that had a major imprint on crustal reworking that shaped the fundamental
1841 architecture of the oldest Archean cratonic core in the Western Dharwar Craton ([Bouhallier et](#)
1842 [al., 1993, 1995](#)).

1843

1844 *9.5 Late granodiorite to granite injections (3150-3100 Ma)*

1845 Late granodiorite to granite injections post-dates previously discussed granitoids and are
1846 commonly found as veins or dykes cross cutting the main foliation patterns in virtually all
1847 granitoid outcrops. In addition, large elongated sheets of north-south direction and coherent
1848 with north-south trending shear zones affected the Eastern block (Bouhallier et al., 1993). They
1849 contain quartz, abundant plagioclase (An₁₂₋₁₄), K-feldspar, magmatic epidote, hornblende,
1850 biotite, zircon, titanite and Fe-Ti oxides. U-Pb zircon ages indicate 3145 -3103 Ma (Jayananda
1851 et al., 2015) coinciding with the metamorphism (Jayananda et al., 2013; Dasgupta et al., 2019).
1852 Their elemental characteristics such as high SiO₂ (SiO₂-76.5-77.0 wt.%), but very low to high
1853 K₂O- 0.34-3.84 wt.% coupled with Nd isotope ($\epsilon_{Nd(T)}$ = 0 to -0.3 Jayananda et al., 2015) and
1854 magmatic epidote suggest rapid injection of magmas originating from a chondritic to depleted
1855 juvenile lower crustal source with possible minor ancient crustal contamination (Jayananda et
1856 al., 2015).. The injection of these granites mark regional metamorphism and terminal event
1857 of the cratonic core formation.

1858

1859 *9.6 Granitoids around other greenstone belts in the cratonic core*

1860 Our data together with published record (Bhaskar Rao et al., 1983; Naqvi et al 1983; Callahan
1861 and Rogers, 1987; Chardon, 1997; Jayananda et al., 2019; Sebastian et al., 2021) for gneisses
1862 and other granitoid rocks adjacent to the Nuggihalli, Kalyadi, J.C. Pura, Chitradurga belt, and
1863 Nagamangala greenstone belt contain quartz, plagioclase (An₁₂₋₂₇), minor K-feldspar, biotite,
1864 hornblende with accessory zircon, apatite, titanite and opaque phases. They show the same
1865 geochemical features as high-Al granitoids (Fig. 7.10k-l, 9a-f) which, therefore, point to a
1866 similar petrogenetic processes and possibly genetic link with the 3400-3300 Ma granitoids
1867 around Holenarsipur greenstone belt.

1868

1869 **10. Geodynamic implications**

1870 *10.1 Crust-mantle differentiation*

1871 Incompatible elements (Th, U, Nb, Zr, Hf, REE) together with Nd-Hf isotope data of Archean
1872 komatiites and basalts reveal that mantle reservoirs from which komatiite magmas originated
1873 were already depleted (Jahn et al., 1982; Jochum et al., 1991; Blichert-Toft and Arndt, 1999;
1874 [Condie, 2003](#); [Chavagnac, 2004](#); [Campbell, 2003](#); [Kerrich and Xie, 2002](#); [Polat et al., 2000](#);
1875 [Mole et al 2015](#); [Jayananda et al., 2008, 2016](#)). Nd isotopic data from the present study
1876 combined with published record (see **fig. 7.11b**) on the 3400-3200 Ma komatiites and
1877 komatiitic basalts ([Jayananda et al., 2008](#); [Mukherjee et al., 2012](#); [Maya, et al.,2017](#); [Patra et](#)
1878 [al., 2020](#); [Ravindran et al., 2021](#)) reveal a widespread ultramafic volcanism which originated
1879 from deep mantle reservoirs and that contributed to large scale crustal growth during 3400-
1880 3200 Ma. Incompatible elemental ratios (Nb/U, Nb/Th, Th/Yb, Nb/Yb see **figs.8a-d**), sub-
1881 chondritic to moderate REE contents and Nd isotope data of the komatiites and komatiitic
1882 basalts reveal large-scale mantle heterogeneity (primitive to depleted mantle) which
1883 experienced long-term depletion history. The existence of such 3400-3200 Ma depleted deep
1884 mantle reservoirs in turn imply an earlier episode of extraction of ultramafic magmas or link to
1885 global differentiation of silicate Earth close to 4500 Ma (Boyet and Carlson, 2005).

1886 Our new U-Pb zircon ages for the granitoids coupled with recent published zircon ages
1887 ([Guitreau et al., 2017](#); [Jayananda et al., 2015,2019](#); [Ranjan et al., 2020](#); [Ao et al., 2021](#)) show
1888 major episodes of TTG-type granitoids formation during 3450-3400 Ma and 3350-3270 Ma
1889 time windows, hence, contributing to high rates of continental growth with earlier events of
1890 purported crustal growth at 3500 Ma and 3600 Ma ([Meen et al., 1992](#); [Guitreau et al., 2017](#);
1891 [Ao et al., 2021](#)). Neodymium and Hafnium isotope data are consistent with the existence of
1892 depleted shallow upper mantle reservoirs with very minor or no inputs from pre-existing crust

1893 during formation of the granitoids (TTG-gneisses). These granitoids therefore represent
1894 successive stages of juvenile continental growth of the WDC. Elemental and Nd-Hf isotopic
1895 characteristics of the granitoids (TTG-type) and greenstone volcanic assemblages indicate
1896 large scale chemical and isotopic heterogeneity with long-term depletion history of the
1897 Paleoproterozoic mantle reservoirs from which the preserved crust of the cratonic core was
1898 extracted. The existence of such shallow and deep depleted mantle reservoirs imply earlier
1899 episodes of mantle differentiation through extraction of granitoid precursors and ultramafic-
1900 mafic greenstone volcanic assemblages. Published U-Pb ages and Lu-Hf isotope data for
1901 magmatic and detrital zircons (Nutman et al., 1992; Bhaskar Rao et al., 2008; Hokada et al.,
1902 2013; Lancaster et al., 2015; Guitreau et al., 2017; Ao et al., 2021) together with Nd isotope
1903 [($\epsilon\text{Nd}_{(T)=3350\text{ Ma}}$) = -2.8 with TDM age of 3890 Ma] signatures (Jayananda et al., 2015) and
1904 radiogenic Pb in feldspars of the 3400 Ma old gneisses (Meen et al., 1992) can be attributed
1905 to earlier episodes of mantle differentiation and crustal growth that possibly occurred in the
1906 3600-3800 Ma time window. The above arguments show that the mantle source reservoirs of
1907 the cratonic core in the WDC had a long term depletion history (3800-3200 Ma). The above
1908 lines of arguments suggest extraction of crust in successive stages contributing to major
1909 continental growth and evolution of complementary depleted mantle reservoirs.

1910

1911 *10.2 Geodynamic context of cratonic core formation*

1912 Despite the fact that numerous studies addressed the tectonic context of Archean continental
1913 growth, this topic remains a subject of much debate and discussion after two decades (e.g.,
1914 Bouhallier et al., 1993; Choukroune et al., 1995; Chardon et al., 1996; Polat et al., 2002; Bedard,
1915 2018; Wyman, 2018; Kusky et al., 2018; Brown et al., 2020; Hyung and Jacobson, 2020;
1916 Sotiriou and Polat, 2020; Nutman et al., 2021; Windley et al., 2021; Kusky et al., 2021 and
1917 Sotiriou et al., 2022). Tectonic regimes operating in the Archean Earth were necessarily

1918 controlled by thermal and mechanical state of the lithosphere which, in turn, was controlled by
1919 a hotter mantle (Choukroune et al., 1994; Monteux et al., 2020). Archean cratons differ from
1920 post-archean shields by the existence of unique lithological assemblages that comprise
1921 komatiites in greenstone belts and polyphase grey gneisses with TTG-affinity in granitoids,
1922 hence, suggesting that specific tectonic processes operated during the Archean (Choukroune et
1923 al., 1997). Several models have been proposed for the formation of the Western Dharwar
1924 Craton which involve both vertical motions associated with mantle plume (Choukroune et al.,
1925 1995; Jayananda et al., 2016), horizontal tectonics (Chadwick et al 2000) and combined plume
1926 -arc formation model (Jayananda et al., 2008, 2018). The different lithological associations in
1927 individual greenstone units combined with their elemental and isotope tracers reveal that each
1928 greenstone belt in the cratonic core represent a distinct tectonic unit and it is proposed that the
1929 three distinct tectonic units of the Holenarsipur belt correspond to an oceanic plateau, an
1930 oceanic arc and a preserved section of oceanic crust close to a spreading center (similar to
1931 modern ocean ridge?). Among the three tectonic units of the Holenarsipur greenstone units, the
1932 southwestern block with komatiite-komatiitic basalt and adjacent shallow water sedimentary
1933 sequences like conglomerate- quartzite -pelite (garnet-staurolite-kyanite-muscovite-biotite)
1934 points to an oceanic plateau which is consistent with their elemental and Nd isotope signatures
1935 of their origin from deep depleted to primitive mantle source. The rock assemblages of the
1936 northcentral block like abundant komatiite and komatiitic basalt-basalt-dacite sequence
1937 interlayered with chloritoid-chlorite-garnet bearing pelites imply an oceanic arc possibly
1938 adjacent to a volcanic plateau. The Nd isotope data (chondritic to depleted source) coupled
1939 with the incompatible element signatures (see fig. 8e) of the volcanic rocks of the northcentral
1940 block is also consistent with an oceanic arc setting. In contrast the lithological sequences such
1941 as pillowed basalts, sheeted dykes, gabbros-foliated gabbros, plagiogranite, layered gabbros
1942 and peridotite in the eastern block represents a section of preserved oceanic crust of the

1943 spreading centre with N-MORB and E-MORB signatures as revealed by elemental and Nd
1944 isotope data. Among other stratigraphically equivalent greenstone belts, the lithological
1945 association (komatiite-komatiitic basalt peridotitic to dunite with chromite layers, komatiitic
1946 basalt-gabbro-anorthosite-plagiogranite) coupled with Nd isotope and elemental signatures (N-
1947 MORB to E-MORB) of the Nuggihalli-Kalyadi belts could also correspond to a preserved
1948 section of oceanic crust. On the contrary, thick and voluminous komatiite-komatiitic basalts
1949 with minor high-Mg basalts together with elemental and Nd isotope tracers showing dominant
1950 primitive to depleted mantle signatures of the J.C. Pura, Banasandra, Ghattihosahalli and
1951 Nagamangala greenstone belts suggest a volcanic plateau setting (Jayananda et al., 2008, 2016;
1952 Tushipokla and Jayananda, 2013). To summarize, the preserved greenstone assemblages in the
1953 cratonic core of the WDC probably represent preserved oceanic volcanic plateaus, oceanic arcs
1954 and oceanic ridge formed within time window of 3400-3200 Ma.

1955 The identification of the three distinct greenstone assemblages in the Holenarsipur
1956 greenstone belt, and equivalents in the cratonic core, with independent tectonic histories is a
1957 very important discovery because their assembly into cratonic framework requires some sort
1958 of horizontal motion during the Paleoproterozoic. In addition, the granitoids (TTG-type gneisses)
1959 adjoining to the Holenarsipur greenstone belt exhibit coeval ages of 3430 Ma to 3300 Ma in
1960 the SW block, whereas 3430-3339 Ma adjoining to Northcentral block, and 3285-3276 Ma to
1961 the western boundary of the eastern block. Strain fabrics data (Bouhallier et al., 1993) and U-
1962 Pb zircon ages indicate that assembly of the three tectonic units (greenstone and adjoining
1963 granitoid gneisses) coincide with the intrusion of the 3230-3177 Ma diapiric trondhjemites,
1964 therefore, suggesting that pre-3200 Ma horizontal motion terminated with the assembly of
1965 different tectonic units (oceanic plateaus, arcs, intervening continental fragments and ocean
1966 ridge crust) and intrusion of trondhjemites plutons. These field characteristics and timing are
1967 shared by granitoids and detrital zircons from the adjoining greenstone belts in the cratonic

1968 core (Meen et al., 1992; Jayananda et al., 2008, 2019; Bidyananda et al., 2016; Ao et al., 2021;
1969 Ranjan et al., 2022).

1970 There is consensus on the fact that Archean granitoids and especially the TTGs formed by
1971 hydrous partial melting of young basaltic precursors at different depths in pressure-
1972 temperature conditions where amphibole and garnet (\pm plagioclase, \pm ilmenite) are stable in
1973 melting residues (Moyen and Martin, 2012; Nédélec et al., 2012). These conditions are
1974 necessary to account for the mineralogy and geochemistry of the TTG-type grey gneisses to
1975 which the WDC granitoids belong. However, the geological and tectonic process that allows
1976 these conditions to be reached is still a matter of debate (e.g., Smithies et al., 2009; Johnson
1977 et al., 2017; Bédard, 2018; Moyen and Laurent, 2018; Hildebrand et al., 2018; Hastie and
1978 Fitton, 2019). The ideal candidate among known, and clearly identified geological process is
1979 the subduction because it can bring material down into the mantle depth at rates fast enough
1980 to avoid complete dehydration of the oceanic slab before reaching the hydrous solidus of wet
1981 basalt. However, some workers (e.g., Rey and Coltice, 2008) argue that rheological
1982 parameters of the Archean crust does not seem to be in favor of plate tectonics in the
1983 Paleoproterozoic. In contrast several other workers argue for horizontal motion of tectonic plates
1984 with eventual subduction as early as Eoarchean (Kusky et al., 2013; Hynes, 2014; Hastie and
1985 Fitton, 2019; Korenaga, 2021; Windley et al., 2021; Nutman et al., 2021). Alternative
1986 models for plate tectonics does exist which invoke horizontal motion that would result in
1987 either intermittent subduction (Van Hunen and Moyen, 2012), formation of slices that pile
1988 downward (de Wit, 1998), or formation of duplex structures (Komiya et al., 2015). Recent
1989 numerical modeling also propose that some localized horizontal motions can develop at the
1990 margin of massive mantle plumes (e.g., Gerya, 2014; Baes et al., 2016) which could reconcile
1991 many of contradictory observations and fit in a fair amount of the Archean tectonic models
1992 involving both vertical and horizontal motions during the same time frame (e.g., Choukroune

1993 et al., 1997; Jayananda et al., 2008; Guitreau et al., 2012; Nagel et al., 2012; Martin et al.,
1994 2014; Bédard, 2018). Recent phase-equilibrium and mechanical modelling study (Mioceovich
1995 et al., 2022) on the Lewisian gneiss complex concluded that sagduction is not a process
1996 viable in the Archean Earth. However, the small volume of mafic layers present in the high-
1997 grade Lewisian gneiss complex together with the fact that this terrane is small compared to
1998 other well-studied lower-grade Archean cratons (e.g., Pilbara-Western Australia, Barberton-
1999 South Africa, Superior Province-Canada) is probably not sufficient to firmly conclude on to
2000 the viability of sagduction for the entire Archean crust. Yet, their model may be region
2001 specific. Therefore, we favour a model in which large volumes of high-density ultramafic-
2002 mafic volcanics associated with TTG type granitoids in the cratonic core of the western
2003 Dharwar craton possibly caused inverse density stratification during reheating of crust
2004 spatially linked to the emplacement of hot trondhjemite magmas close to ca.3200 Ma, in a
2005 way perhaps akin to the environments described in Choukroune et al. (1995, 1997).
2006 Furthermore, detailed strain fabrics mapping and kinematic analysis across Paleoproterozoic
2007 cratons (e.g., Kaapvaal and Pilbara) also reveal partial convective overturns which likely
2008 drove sagduction processes during the Paleoproterozoic (Collins et al., 1998; Gardiner et al.,
2009 2017; Peterson et al., 2019; Wiemer et al., 2018).

2010 Sagduction is a vertical tectonic process clearly identified in the Archean (e.g., Gorman
2011 et al., 1978; Bouhallier et al., 1995; Chardon et al., 1996) that could possibly replace the need
2012 for horizontal tectonics (e.g., Van Kranendonk et al., 2014). In addition, proposed sources for
2013 the formation of TTG-type granitoids match well with tholeiitic basalts present in greenstone
2014 belts (e.g., Rapp and Watson, 1995; Nédélec et al., 2012; Guitreau et al., 2012; Martin et al.,
2015 2014). As mentioned earlier, the western Dharwar craton presents a continuous tilted crustal
2016 panel from upper to lower crust which, hence, allows the sinking of the high-density greenstone
2017 volcanics of the Dharwar craton to be followed. Yet, sinking greenstone volcanics exhibit a

2018 continuous increase in metamorphic grade until granulite facies (e.g., [Raase et al., 1986](#);
2019 [Jayananda et al., 2013a](#)) without any evidence for *in-situ* partial melting and melt migration
2020 upwards. However, numerical modeling suggests that this process can produce anatectic melts
2021 ([Nicoli, 2020](#)) but it would have to be investigated further if geochemical signatures of 3600-
2022 3300 Ma granitoids in the cratonic core of the WDC could be reproduced by such a process
2023 since this author used 3000 Ma granitoids composition to compare to his model whereas field
2024 evidence suggest that the WDC acquired its architecture by 3200 Ma ago (e.g. [Jayananda et](#)
2025 [al., 2018](#)).

2026 Regardless of the capacity of sagduction to induce granitoid formation, sagduction
2027 alone cannot account for the assembly of the three independent tectonic blocks around
2028 Holenarsipur greenstone belt and other tectonic units formed in the same time window in the
2029 cratonic core. Therefore, our preferred model for the formation of the cratonic core in the
2030 western Dharwar craton is as follows. Vertical motions of mantle plumes in upwelling zones
2031 produced oceanic plateaus during 3400-3300 Ma, which possibly initiated horizontal motions
2032 due to boundary forces. Horizontal density contrasts could have facilitated the triggering of
2033 protosubduction (e.g., duplex, downward piling of slices) and, hence, the development of
2034 oceanic arcs, ridges (e.g., [Nair and Chacko, 2008](#)), which lead to generation of the protoliths
2035 for the 3400-3300 Ma granitoids (TTG-type gneisses) thus contributing to major juvenile
2036 continental growth. The location of these horizontal movements could have also been
2037 facilitated by rheological weakness ([Rey and Coltice, 2008](#); [Rey et al., 2014](#)) that could be due
2038 to the presence of pre-existing granitoids such as those older than 3430 Ma and up to 3600 Ma
2039 in adjacent to some of the oceanic plateaus (e.g., [Guitreau et al., 2017](#)). The assembly of
2040 different tectonic units ended with the culmination of some sort of subduction of intervening
2041 oceanic crust between oceanic plateaus/ridges and eventual slab breakoff causing mantle
2042 upwelling which resulted in melting of the newly formed juvenile crust and pre-existing

2043 granitoid crust (3600-3300 Ma) at different levels (14-18 kbar), in turn, forming trondhjemite
2044 magmas during 3200 Ma and their intrusion mark the terminal magmatic event of juvenile
2045 crustal growth in the cratonic core. The hot trondhjemite magma intrusions associated with
2046 the mantle upwelling caused the partial convective overturn of the crust which drove the
2047 gravitational collapse of high-density greenstone material and rise of low-density domes that
2048 resulted in dome-keel patterns as revealed by the strain fabrics mapping and kinematic analysis
2049 on the scale of cratonic core ([Bouhallier et al., 1993,1995](#); [Choukroune et al., 1995](#); [Chardon et](#)
2050 [al., 1996](#)). This dome-keel formation is spatially associated with medium to high grade
2051 metamorphism, crustal reworking and granite emplacement (granite sheets injected into the
2052 gneisses of southwestern block and Chikmagalur granite; [Jayananda et al., 2015](#)) and
2053 cratonization of Paleoproterozoic crust close to 3100 Ma ([Jayananda et al., 2013, 2015](#); [Dasgupta](#)
2054 [et al., 2019](#)). The cratonized mature crustal segment form provenance for the detritus (e.g.,
2055 quartzites and conglomerates) in the <3100 Ma sedimentary basins formed ([Hokada et al.,](#)
2056 [2013](#); [Chatterjee and Das, 2004](#); [Jayananda et al., 2018](#); [Ranjan et al., 2022](#)). Finally, crustal
2057 adjustments through metamorphic cooling, crustal reworking, and 3000 Ma hybrid potassic
2058 granite formation (e.g., Bellur-Nagamangala potassic granites) marked the final cratonization
2059 of the cratonic core in the WDC (e.g., [Jayananda et al., 2019, 2020](#); [Ao et al., 2021](#)).

2060 This model may apply to other parts of the WDC and spatially associated blocks that
2061 recorded similar histories either within the sedimentary record or the meta-igneous record. This
2062 is the case for the Bababudan basin ([Jayananda et al., 2015](#); [Lancaster et al., 2015](#); [Bidyananda](#)
2063 [et al., 2016](#)), western margin of the Chitradurga greenstone belt ([Chardon et al., 1997](#);
2064 [Lancaster et al., 2015](#); [Wang and Santosh, 2019](#); [Ravindran et al., 2021](#)), Coorg massifs
2065 ([Santosh et al., 2015](#); [Roberts and Santosh, 2018](#)), Sakaleshpur and Western Ghats ([Jayananda](#)
2066 [et al in prep.](#)). For instance, zircon U-Pb ages of magmatic and detrital zircons from Coorg
2067 massifs were attributed to three major crust forming events at 3500 Ma, 3370-3270 Ma, and

2068 3190-3140 Ma (Santosh et al., 2015; Amal Dev et al., 2016; Roberts and Santosh, 2018) with
2069 remnants as old as 3750 to 4030 Ma (Santosh et al., 2015). The Coorg massif is considered as
2070 an exotic Mesoarchean supercontinent (Santosh et al., 2015) assembled with the Dharwar
2071 craton during Mesoproterozoic but our zircon age data (3500-3200 Ma) between the
2072 Holenarsipur greenstone belt and Coorg massifs (Jayananda et al unpublished data) together
2073 with a recent study (Roberts and Santosh, 2018), in contrast, suggest a common crustal building
2074 history of 3500-3200 Ma followed by granite intrusion and metamorphism at 3150-3100 Ma
2075 in the context of collision leading to the stabilization of the craton close to 3100 Ma.

2076

2077 *10.3 Comparison with coeval cratons and global implications*

2078 Archean crustal nuclei originated as micro-blocks in diverse tectonic environments assembled
2079 through lateral motions which resulted in large cratonic cores with thick crustal roots. Our
2080 study in the western Dharwar craton reveals that different tectonic units were juxtaposed
2081 through horizontal motion culminating into subduction (3350-3230 Ma) with eventual slab
2082 break off drive mantle upwelling causing melting of thickened crust resulting in generation of
2083 the trondhjemite magmas. The intrusion of these trondhjemite magmas, in turn, caused partial
2084 convective overturn which led to the development of dome-keel structures at 3200 Ma,
2085 followed by HT/HP metamorphism, late-kinematic minor granite intrusions during 3145-3100
2086 Ma marking the final stabilization of cratonal core with the development of crustal roots which
2087 supplemented by seismic tomographic data (Gupta et al., 2003) indicating thick crust (45-55
2088 km).

2089 The 3600-3100 Ma crustal history of the western Dharwar craton correlates with the
2090 crustal growth history and thermal record of the Singhbhum craton, Eastern India (Upadhyay
2091 et al., 2014), the Kaapvaal craton, South Africa (Kroner et al., 2014), Pilbara craton, western
2092 Australia (Van Kronendonk et al.2015), Sao Francisco craton (Teixeira et al., 2017), the Slave

2093 province ([Bleeker 2002](#)). The TTG-greenstone assemblages in the Singhbhum craton preserve
2094 a record of two major stages of magmatism including 3450-3440 Ma tonalitic to trondhjemite
2095 and 3350-3320 Ma granodiorite to granite phase ([Upadhyay et al., 2014](#)) with vestiges of TTG-
2096 greenstone assemblages as old as 3610-3507 Ma ([Mukhopadhyay et al., 2008](#); [Nelson et al.,](#)
2097 [2014](#)). These Paleoarchean crust building events were followed by a partial convective
2098 overturn, thermal event, and craton formation at 3200-3100 Ma, hence, comparable to the
2099 crustal history of the cratonic core in the western Dharwar craton.

2100 The Kaapvaal craton contains microblocks which comprise greenstone belts and TTGs
2101 formed during 3600-3300 Ma and assembled through lateral motion in the context of
2102 subduction, thereby causing thickening, metamorphism, reworking, deformation, and
2103 generation of granites ca 3230-3100 Ma ([Kamo and Davis, 1994](#); [Schoene et al., 2008](#)). Both
2104 the Holenarsipur and the Barberton greenstone belts with their adjacent granitoids of TTG-
2105 affinity and intrusive plutons share comparable protracted crustal histories in term of
2106 lithological assemblages, magmatism, dome-keel patterns, and metamorphic record ([Compston](#)
2107 [and Kroner, 1988](#); [Van Kranendonk et al., 2010](#); [Jayananda et al., 2018](#); [Wang et al., 2019](#)).
2108 The Barberton greenstone belt is an assembly of allochthonous sequences with independent
2109 protracted crustal and tectonic histories which were interpreted as juxtaposed through active
2110 convergence ([Schoene et al., 2008](#)). The volcanic sequences from different stratigraphic units
2111 of the Barberton greenstone belt yield ages of 3644 to 3300 Ma ([Armstrong et al., 1990](#); [Kroner](#)
2112 [et al., 1996](#); [Byerly et al. 1996](#); [Chavagnac, 2004](#)) whereas the surrounding granitoids (TTG-
2113 type gneisses) extending into Swaziland formed in major events between 3660 and 3300 Ma,
2114 and granite plutons emplaced at 3230-3100 Ma ([Compston and Kröner, 1988](#); [Kroner et al.,](#)
2115 [2014](#); [Zeh et al., 2011](#); [Armstrong et al., 1990](#); [Kamo and Davis, 1994](#)). The intrusion of hot
2116 trondhjemite magmas into the greenstone sequence and adjoining granitoids (TTGs) at 3226-
2117 3180 Ma initiated gravity driven partial convective overturn wherein colder high-density

2118 greenstones collapse into low density softer granitoids (TTG-type gneisses) resulting in dome-
2119 keel structures ([Van Kranendonk et al., 2010](#)). The 3200 Ma old garnet-bearing high-pressure
2120 metamorphic assemblages in the Barberton region interpreted to be resulted by partial
2121 convective overturn during formation of plutons ([Wang et al., 2019](#)). On the contrary, the
2122 Assegaai De Kraalan and Witrivier greenstone belts from the southeast of the Kaapvaal craton
2123 contain dominant ultramafic-mafic volcanic rocks with intercalated chert and BIFs. These
2124 greenstone sequences are dated at 3222 ± 8 Ma and 3193 ± 5 Ma intrusions and show an initial
2125 low T/P followed by high T/P metamorphic assemblages which is overall interpreted as a
2126 response to crustal thickening through convergence during Paleoproterozoic ([Saha et al., 2010](#)).

2127 The east Pilbara craton in Western Australia preserves an Archean crustal record with granite-
2128 greenstone assemblages display classical dome and basin patterns which share similarities with
2129 the oldest cratonic core (3600-3200 Ma) of the western Dharwar craton. The Archean
2130 continental nuclei in the east Pilbara craton built in successive stages through granitoid
2131 formation and greenstone volcanism during 3660-3220 Ma ([Nelson et al., 1999](#); [van](#)
2132 [Kranendonk et al., 2015](#); [Gardiner et al., 2017](#); [Peterson et al., 2019](#); [Wiemer et al., 2018](#); [Kusky](#)
2133 [et al., 2021](#)). The oldest crustal remnants identified are dated from 3660 to 3580 Ma using
2134 zircon in tonalitic gneiss enclaves and metagabbro whereas detrital U-Pb zircon ages reveal
2135 3710 Ma ([Thorpe et al., 1992](#)), thus, indicating the existence of an even older crust. U-Pb dating
2136 of zircons from leucogabbro and gabbro indicate 3588 ± 5 Ma and 3578 ± 5 Ma, respectively
2137 ([Peterson et al., 2019](#)) which are correlated with similar U-Pb zircon ages for TTGs from
2138 Muccan dome. Furthermore, 3467 Ma U-Pb zircon age of TTG suite from Warrawagine dome
2139 ([Kemp et al., 2017](#)), and U-Pb zircon ages of 3484-3462 Ma from Callina Supersuite, 3451-
2140 3416 Ma Tambina Supersuite and 3324-3277 Ma from Emu Pool Supersuite and 3274-3223
2141 Ma Cleland Supersuite ([Van Kranendonk et al., 2007](#); [Gardiner et al., 2017](#); [Peterson et al.,](#)
2142 [2019](#)) are also consistent. The TTGs suite formation terminated with younger 3250-3220 Ma

2143 magmatic event of monzogranite and syenogranite formation which belong to Cleland
2144 Supersuite (Gardiner et al., 2017; Peterson et al., 2019). To summarize, widespread granitoid
2145 magmas with TTG affinity emplaced during 3600-3300 Ma and contemporaneous greenstone
2146 volcanism contributed to the continental growth (Nelson et al., 1999; Van Kranendonk et al.,
2147 2007; Gardiner et al., 2017; Peterson et al., 2019) followed by partial convective overturn of
2148 crust which resulted in dome and keel patterns supposedly developed at 3325-3300 Ma (Collins
2149 et al., 1998) and followed by late magmatic intrusions ca.3250-3200 Ma (Gardiner et al., 2017;
2150 Peterson et al., 2019; Wiemer et al., 2018). On the contrary a more recent study (Kusky et al.,
2151 2021) argues that development of large domal structures is linked to late orogenic shortening
2152 which was synchronous with or soon-after intrusion of 3274-3223 Ma plutons.

2153 The Archean crustal record (3650-3180 Ma) of the northern Sao Francisco craton
2154 (Teixeira et al., 2017) in Brazil correlates with the formation history of the cratonic core in the
2155 WDC. The preserved Archean crust (TTG-greenstone assemblages and intrusive granites)
2156 formed through a multi-stage history. U-Pb zircon ages reveal that the oldest TTGs in Roacho
2157 de Santana formed at 3648 ± 69 Ma whereas remnants of the Hadean record were found as
2158 detrital zircon (4100 Ma; Paquette et al., 2015). Widespread TTG formation at 3403-3325 Ma
2159 (Nutman and Cordani, 1993; Martin et al., 1997; Dantas, et al., 2013; Santos-Pintos et al., 2012)
2160 and 3305 ± 9 Ma greenstone volcanism (Peucat et al., 1995) contributed to large scale
2161 continental growth. Late-stage granodioritic to granite intrusions dated at 3250 Ma (Bastos
2162 Leal et al., 2003), 3184 ± 6 Ma (Nutman and Cordani, 1993), 3158 ± 2 Ma (Martin et al., 1997)
2163 and 3040 Ma (Peucat et al., 2002) represent terminal juvenile additions followed by reworking
2164 of ancient crust (Barbosa et al., 2013) and stabilization of the Sao Francisco craton.

2165 Globally, the 3600-3100 Ma time window appears to represent a major period of crust-
2166 mantle differentiation, continental growth, reworking of pre-existing crust, LP-HT
2167 metamorphism, and eventual formation of stable Archean cratonic cores. The continental

2168 growth and craton formation involved episodic formation of granitoids and coeval, though
2169 unrelated, greenstone volcanic rocks early in mantle upwelling zones followed by arc setting
2170 resulting in the formation of diachronous blocks representing cycles of magmatism and
2171 sedimentation. Assembly of these crustal blocks was likely driven through horizontal motions
2172 with eventual convergence in arc settings and slab breakoff leading to high pressure melting of
2173 mafic crust and generation hot-felsic magmas and hybrid granitoids. Formation of these hot
2174 magmas induced partial convective overturn of rheologically weaker low-density felsic crust.
2175 This led to a gravitational collapse of high-density greenstones into low density infracrustal
2176 gneisses and the development of regional dome-keel patterns, HT/HP metamorphism and
2177 cratonization at 3200-3100 Ma. Small differences observed between Archean cratons can
2178 certainly arise from distinct abundances of involved lithological units, as well as local age
2179 differences and pre-histories. Furthermore, Windley et al., (2021) suggested that the onset of
2180 horizontal motion of tectonic plates began by the Eoarchean and contributed to the Archean
2181 continental growth through collision of arcs and continents whilst the development of domal
2182 patterns in Archean cratons are attributed to orogenic shortening and related deformation
2183 caused by late magmatic intrusions originated during massive slab breakoffs in subduction
2184 zones (Kusky et al., 2021). Long-term accumulation of evidence suggests that the Earth's
2185 global geodynamics evolved throughout the Archean to progressively initiate modern-plate
2186 tectonics (e.g. [Korenaga, 2013](#); [Moyen and Laurent, 2018](#)) which did not likely occurred
2187 everywhere at the same time but most likely gradually evolved from local and punctual tectonic
2188 events to a more globally-connected system of plates (e.g. [Laurent et al., 2014](#)). However, it is
2189 important to note that the horizontal motions we have evidenced in the Western Dharwar
2190 craton, and which we compare to other cratons, are valid regardless of their actual origin (e.g.,
2191 plate tectonics, lower crustal flow, plume-related lithospheric adjustment).

2192

2193 **11. Conclusions**

2194 This synthesis integrates our new data with published record of field, petrographic,
2195 geochronologic, elemental and isotope (Nd-Hf-Sr-Pb) data of greenstone volcanic rocks,
2196 granitoids (TTG-type) and diapiric trondhjemites from the cratonic core of the western
2197 Dharwar craton and address processes involved in the origin of the oldest continental nuclei,
2198 evolving tectonics of the Archean earth and formation of the cratonic cores worldwide. The
2199 salient conclusions of this synthesis can be summarized as follows:

- 2200 1. The oldest preserved crustal nuclei in the western Dharwar craton comprises komatiite-
2201 dominated greenstone assemblages, TTG-type grey gneisses and diapiric trondhjemites
2202 forming a wide time (3600-3200 Ma) and tectonic window for the early earth dynamics,
2203 crust-mantle evolution and formation of cratonic cores.
- 2204 2. The preserved crust in the cratonic core accreted in three major episodes during 3430-
2205 3400 Ma, 3350-3270 Ma and 3230-3170 Ma with remnants of crust formed between
2206 3500 and 3600 Ma.
- 2207 3. Field evidence together with elemental and Nd isotope signatures of greenstone
2208 volcanic assemblages in the cratonic core suggest their independent histories which
2209 point to an origin in primitive to depleted mantle reservoirs probably formed as oceanic
2210 plateaus, oceanic arc and oceanic crust section of the spreading center.
- 2211 4. The magmatic precursors of the TTG-type granitoids originated by partial melting of
2212 short-lived mafic crust at different depths (10-15 kbar) without or with variable garnet
2213 in source residue whilst diapiric trondhjemite magmas formed by melting of base of the
2214 (thickened) mafic- tonalitic crust with probable garnet in residue.
- 2215 5. Horizontal motion of the oceanic crust intervening to pre-existing continental nuclei,
2216 oceanic plateaus, arcs and ridges during 3200 Ma led to the assembly of different
2217 tectonic elements into the cratonic core. Continued subduction with slab breakoff

2218 caused asthenosphere upwelling that caused melting of thickened crust generating
2219 trondhjemite magmas which upon upward movement caused partial convective
2220 overturn and gravitational collapse of high-density greenstones into low-density
2221 granitoids of TTG-affinity causing dome-keel structures and HT-LP metamorphism.
2222 6. The crust building processes, continental growth and craton forming mechanisms
2223 documented in the cratonic core of the WDC also documented in other Archean cratonic
2224 cores in Peninsular India, Southern Africa, Western Australia and South America.
2225 Therefore it appears that globally 3600-3200 Ma time window corresponds to major
2226 period of mantle differentiation, high rates of juvenile crustal growth, transient tectonics
2227 and craton formation worldwide.

2228

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2243

2244 **Author contributions**

2245 M.J designed the project, carried out field work, generated and compiled the data base and
2246 wrote the original manuscript; M.G participated in field work, generated data related to zircon
2247 (i.e., images, U-Pb, and Lu-Hf) , co-drafted some figures, provided intellectual inputs in data
2248 interpretation, and participated in the writing of the present manuscript. KRA participated in
2249 field work, compiling the data, drafted figures and provided intellectual inputs; T.M and SLC
2250 have analyzed greenstone volcanic rocks for elemental and Nd isotopes and provided
2251 intellectual inputs.

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2253

2254

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3377 **Figure captions**

3378 **Fig.1-** Geological sketch map of Peninsular India showing major Archean cratons in India
3379 (modified after [Geological Survey of India](#)), square box depicting the cratonic core region.

3380 **Fig.2 -** Geological sketch map of Archean Dharwar craton and study area (modified after
3381 [Chardon et al., 2008, 2011; Peucat et al., 2013; Jayananda et al., 2013a](#)). (a) Geological sketch
3382 map showing the cratonic core (b) Geological sketch map of Holenarsipur greenstone belt and
3383 adjoining TTGs showing three greenstone units and diapiric trondhjemite intrusions with age,
3384 ϵHf , ϵNd of the present study; [Jayananda et al., 2015; Guitreau et al., 2017; Ranjan et al., 2020;](#)
3385 [Ao et al., 2021](#), foliation trajectories are from [Bouhallier et al., 1993](#) (c) SW-NE interpretative
3386 cross section of crustal panel of Holenarsipur greenstone belt and adjoining felsic rocks with
3387 distinct lithological assemblages corresponding to three tectonic blocks (modified after
3388 [Bouhallier et al., 1995](#)) and this interpretative cross-section represents the A-A' shown in the
3389 fig. 2b.

3390 **Fig.3.1 (a)** Komatiitic basalt interlayered with grey gneiss; **(b)** Komatiite with pillow structure
3391 **(c)** Tattékere conglomerate; **(d)** ripple marks in quartzite; **(e)** cross-bedding in quartzite; **(f)**
3392 Kyanite-staurolite-garnet bearing pelite **(g)** light grey granite with whitish grey trondhjemite in
3393 Hassan-Gorur road; **(h)** grey granodiorite with older gneissic enclave in Hassan-Gorur road; **(i)**
3394 late (3145 Ma) granodiorite dyke injected into TTG in Gorur-Hassan road; **(j)** Trondhjemite
3395 forming vertical tectonite close to Karle.

3396 **Fig. 3.2 (a)** Komatiite showing pillow structure from Northcentral block about 1km east of
3397 Haradanahalli; **(b)** Pyroclastic felsic volcanic flow at Eastern outskirts of Haradanahalli; **(c)**
3398 Chlorite-chloritoid-garnet bearing pelitic rock at about 1 km east of Shigaranahalli; **(d)**
3399 Randomly oriented crystals of actinolite, tremolite and hornblende at about 500m NE of
3400 Haradanahalli.

3401 [Fig. 3.3 \(a\)](#) Eastern block road cut section displaying section of preserved oceanic crust from
3402 pillow ultramafic through clays, BIFs, sheeted dykes, plagiogranite, gabbro/norite, layered
3403 gabbro with ultramafic and finally peridotite and this corresponds to the B-B' line shown in
3404 the fig.2b; [\(b\)](#) Interpretative E-W section of preserved oceanic crust. [\(c\)](#) Interpretative vertical
3405 section of the oceanic crust.

3406 [Fig. 4.1 \(a\)](#) Photomicrograph from southwestern block [\(b & c\)](#) from northcentral block [\(d\)](#) from
3407 eastern block.

3408 [Fig. 5.1 \(a\)](#) The Nuggihalli greenstone belt showing metamorphosed ultramafic rocks
3409 (peridotite, serpentinite and dunite with chromite layers, talc-tremolite schists [\(b\)](#) the Kalyadi
3410 greenstone belt showing komatiite volcanic rocks including serpentinites, serpentine-talc-
3411 tremolite and talc-chlorite schists [\(c & d\)](#) the J.C. Pura greenstone belt showing komatiite-
3412 komatiitic basaltkomatiitic basalts and minor basalt with flow top pillow breccias [\(e\)](#) the
3413 Banasandra greenstone belt showing spinifex texture near Kunikenahalli [\(f\)](#) the Nagamangala
3414 greenstone belt showing komatiite affinity with minor basalts which are interlayered with
3415 sedimentary rocks.

3416 [Fig. 5.2 \(a\)](#) Photomicrograph from the Kalyadi greenstone belt [\(b\)](#) the Nuggihalli greenstone
3417 belt [\(c\)](#) the J.C. Pura greenstone belt [\(d\)](#) the Ghattihosahalli greenstone belt [\(e\)](#) the Banasandra
3418 greenstone belt [\(f\)](#) the Nagamangala greenstone belt.

3419 [Fig. 6.1 \(a\)](#) TTG-type granitoids at the base of Bababudan [\(b\)](#) TTG-type granitoids west of the
3420 Nuggihalli greenstone belt [\(c\)](#) TTG-type granitoids west of the Chitradurga greenstone belt
3421 near Bharamanayakana durga [\(d\)](#) TTG-type granitoids on the Huliyaar-Hosdurga road [\(e\)](#) TTG-
3422 type granitoids near the J.C. Pura [\(f\)](#) TTG-type granitoids west of the Nagamangala greenstone
3423 belt. [\(g\)](#) Interpretative WSW-ENE crustal scale cross- section of the cratonic core in WDC
3424 based on our field observations and published tectonic fabrics data ([Bouhallier et al., 1993](#),

3425 1995; Chardon, 1997; Chardon et al., 1996, 1998, 2008; the Moho depth is adopted from Gupta
3426 et al., 2003).

3427 Fig. 7.1 (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
3428 al., 1982) showing that volcanic rocks of three greenstone unit are komatiite to komatiitic basalt
3429 composition (b) $\text{CaO-MgO-Al}_2\text{O}_3$ ternary plot (Viljoen et al., 1982) showing dominant
3430 komatiite to komatiitic basalt except two samples of Eastern block showing tholeiite
3431 composition. (c) $\text{Feo}/(\text{Feo}+\text{MgO})$ versus Al_2O_3 (Arndt, 2003) binary diagram showing
3432 Barberton komatiite and komatiitic basalt field (d) Harker's binary variation diagram for major
3433 oxide against MgO showing HGB and other greenstones from cratonic core region (e) Harker's
3434 binary variation diagram for trace element against MgO showing HGB and other greenstones
3435 from cratonic core region.

3436 Fig. 7.2 (a) Chondrite normalized low total REE group volcanic rocks of southwestern block
3437 (b) Primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of low total
3438 REE group southwestern block (c) Chondrite normalized high total REE group volcanic rocks
3439 of southwestern block (d) Primitive mantle normalized (Sun and McDonough, 1989) multi-
3440 element plot of high total REE group of southwestern block (e) Chondrite normalized REE
3441 group of komatiite volcanic rocks of northcentral block (f) Primitive mantle normalized (Sun
3442 and McDonough, 1989) multi-element plot of komatiite volcanic rocks of northcentral block.
3443 (g) Chondrite normalized REE group of komatiitic-basalts of northcentral block (h) Primitive
3444 mantle normalized (Sun and McDonough, 1989) multi-element plot of komatiite-basalts of
3445 northcentral block (i) Chondrite normalized low total REE group volcanic rocks of eastern
3446 block (j) Primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of low
3447 total REE group of eastern block (k) Chondrite normalized high total REE group volcanic rocks
3448 of eastern block (l) Primitive mantle normalized (Sun and McDonough, 1989) multi-element
3449 plot of high total REE group of eastern block.

3450 Fig. 7.3 (a) Chondrite normalized REE group volcanic rocks of the Nuggihalli and Kalyadi (b)
3451 Primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of the Nuggihalli
3452 and Kalyadi.

3453 Fig. 7.4 (a) Chondrite normalized REE volcanic rocks of the J.C. Pura komatiites (b) Primitive
3454 mantle normalized (Sun and McDonough, 1989) multi-element plot of the J.C. Pura komatiites
3455 (c) Chondrite normalized REE volcanic rocks of the J.C. Pura komatiites-cumulate layers (d)
3456 Primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of the J.C. Pura
3457 komatiites-cumulate layers (e) Chondrite normalized REE volcanic rocks of the J.C. Pura
3458 komatiite-basalt (f) Primitive mantle normalized (Sun and McDonough, 1989) multi-element
3459 plot of the J.C. Pura komatiite-basalt.

3460 Fig. 7.5 (a) Chondrite normalized REE volcanic rocks of the Ghattihosahalli komatiites (b)
3461 Primitive mantle normalized (Sun and McDonough, 1989) multi-element plot of the
3462 Ghattihosahalli komatiites (c) Chondrite normalized REE volcanic rocks of the Banasandra
3463 komatiites (d) Primitive mantle normalized (Sun and McDonough, 1989) multi-element plot
3464 of the Banasandra komatiites (e) Chondrite normalized REE volcanic rocks of the
3465 Nagamangala komatiites (f) Primitive mantle normalized (Sun and McDonough, 1989) multi-
3466 element plot of the Nagamangala komatiites.

3467 Fig. 7.6 (a) $(Gd/Yb)_N$ versus Al_2O_3/TiO_2 (after Arndt 2003) (b) $(Gd/Yb)_N$ versus CaO/Al_2O_3
3468 (after Jahn et al. 1982) indicating role of variable residual garnet or absence of residual garnet
3469 in source mantle.

3470 Fig. 7.7 - (a) Cathodoluminescence (CL) images of the sample 18HP01A, 18HP01C,
3471 18HP03A, 18HP03B and 18HP04A (b) cathodoluminescence (CL) images of the sample
3472 18HP04A, 18HP04B, 18HP05, 18HP07 and 18HP10A (c) cathodoluminescence (CL) images
3473 of the sample 18HP13, 18HP15A, 18HP23, 18HP24A and 18HP39.

3474 [Fig. 7.8](#) (a) U-Pb Concordia diagram of sample 18HP23 southwestern block TTG (b) U-Pb
3475 Concordia diagram of sample 18HP24a southwestern block TTG (c) U-Pb Concordia diagram
3476 of sample 18HP10b southwestern block TTG (d) U-Pb Concordia diagram of sample 18HP05
3477 southwestern block TTG (e) U-Pb Concordia diagram of sample 18HP01a southwestern block
3478 TTG (f) U-Pb Concordia diagram of sample 18HP01c southwestern block TTG (g) U-Pb
3479 Concordia diagram of sample 18HP03a southwestern block TTG (h) U-Pb Concordia diagram
3480 of sample 18HP03b southwestern block TTG (i) U-Pb Concordia diagram of sample 18HP04a
3481 southwestern block TTG (j) U-Pb Concordia diagram of sample 18HP04b southwestern block
3482 TTG (k) U-Pb Concordia diagram of sample 18HP07 northcentral block TTG (l) U-Pb
3483 Concordia diagram of sample 18HP15a northcentral block TTG (m) U-Pb Concordia diagram
3484 of sample 18HP13 felsic volcanic, northcentral block TTG (n) U-Pb Concordia diagram of
3485 sample 18HP39 plagiogranite, eastern block.

3486 [Fig.7.9](#) (a) Normative Ab-An-Or triangular diagram ([after O' Connors, 1965](#)) with fields from
3487 [Barker \(1979\)](#) showing trondhjemite, tonalite and granite extending to granodiorite field (b)
3488 normative Q–Ab–Or ([Barker and Arth, 1976](#)) showing trondhjemite and calc-alkaline trend.
3489 (c) Harker's binary diagrams for TTG-type granitoids and diapiric trondhjemite intrusions
3490 displaying moderate to weak fractionation trends.

3491 [Fig.7.10](#) (a) Chondrite normalized low-Al TTG REE group of southwestern block (b) Primitive
3492 mantle normalized ([Sun and McDonough, 1989](#)) multi-element plot of low-Al TTG REE
3493 group of southwestern block (c) Chondrite normalized high-Al TTG REE group of
3494 southwestern block (d) Primitive mantle normalized ([Sun and McDonough, 1989](#)) multi-
3495 element plot of high-Al TTG REE group of southwestern block (e) Chondrite normalized REE
3496 group of TTG from northcentral block (f) Primitive mantle normalized ([Sun and McDonough,](#)
3497 [1989](#)) multi-element plot of TTG from northcentral block. (g) Chondrite normalized REE
3498 group of TTG from eastern block (h) Primitive mantle normalized ([Sun and McDonough,](#)

3499 [1989](#)) multi-element plot of TTG from eastern block (i) Chondrite normalized REE group of
 3500 Trondhjemite (j) Primitive mantle normalized ([Sun and McDonough, 1989](#)) multi-element plot
 3501 of Trondhjemite (k) Chondrite normalized REE group from other western TTG (l) Primitive
 3502 mantle normalized ([Sun and McDonough, 1989](#)) multi-element plot from other western TTG.
 3503 [Fig.7.11 \(a\)](#) $\epsilon\text{Hf}_{(T)}$ versus time evolution diagram of TTGs from three blocks, other western
 3504 TTGs and trondhjemite showing involvement of depleted mantle with traces of ancient crustal
 3505 contamination. (b) $\epsilon\text{Nd}_{(T)}$ versus time evolution diagram of greenstones, TTGs from three
 3506 blocks and other greenstone belts, TTGs in the cratonic core region showing involvement of
 3507 depleted mantle with traces of ancient crustal contamination.
 3508 [Fig.8 \(a\)](#) Nb/U versus Nb/Th plot ([after Saunders et al. 1988](#)) showing scattering indicating
 3509 source composition of primitive mantle to depleted MORB source mantle for the southwestern,
 3510 northcentral, eastern block and other greenstone belts in the cratonic core region (b) Nb/Th
 3511 versus ϵNd plot indicating the primitive mantle and depleted MORB source for the studies
 3512 samples from cratonic core region. (c) Th/Yb versus Nb/Yb binary plot ([Pearce, 2008](#))
 3513 indicating primitive to depleted mantle source for the southwestern, northcentral, eastern block
 3514 and other greenstone belts in the cratonic core region (d) Nb/Y versus Zr/Y ([Condie, 2003](#))
 3515 indicating deep depleted to primitive mantle reservoir source for the samples from for the
 3516 southwestern, northcentral, eastern block and other greenstone belts in the cratonic core region
 3517 (e) Zr/Nb versus Nb/Th plot ([Condie, 2003](#)) showing oceanic plateau for southwestern block,
 3518 J.C.Pura, Ghattihosahalli, majority of Banasandra, Nuggihalli and Kalyadi samples, arc to
 3519 oceanic plateau for Northcentral block, Nagamangala samples and N-MORB to oceanic plateau
 3520 for Eastern block. Arrows indicate effects of batch melting (F) and subduction (SUB); PM,
 3521 primitive mantle; DM, shallow depleted mantle; ARC, arc related basalts; NMORB, normal
 3522 ocean ridge basalt; OIB, oceanic island basalt; DEP, deep depleted mantle; EN, enriched
 3523 component.

3524 Fig.9 (a) Nb/Ta vs Gd/Yb (Hoffmann et al., 2011) showing the origin of TTG melt from
3525 different source composition (b) Nb/Ta vs Zr/Sm plot (Hoffmann et al., 2011) explaining the
3526 origin of TTG melt from different source composition, (c) $K_2O/(Na_2O+CaO)$ vs Sr/Y plot
3527 (Hoffmann et al., 2019) representing the different pressure level at which the studied samples
3528 generated from the source. Lines correspond to melting models derived from experimental
3529 database, as in Moyen, 2009, 2011, at different pressures and for two sources, an MORB and
3530 a more enriched mafic rock (d) $Al_2O_3/(FeO_T+MgO)-3*CaO-5*(K_2O/Na_2O)$ diagram (Laurent
3531 et al., 2014) showing the sources of studies TTGs from the cratonic core region (e) Sr/Y versus
3532 Y diagram (after Defant and Drummond, 1990; Martin et al., 2005) showing two distinct groups
3533 of TTG suites (low-Al and high-Al) and the role of residual garnet. Low-Al gneisses plot in
3534 the field defined for island arc magmas derived from mafic source without significant residual
3535 garnet. High-Al gneisses plot in the field of Archean TTG derived by melting of mafic source
3536 with variable residual garnet. Arrows indicate batch melting of mafic source with 7–30%
3537 garnet-amphibole in source residue (f) $(La/Yb)_N$ versus Yb plot (after Jahn et al., 1981;
3538 Drummond and Defant, 1990; Moyen, 2011) showing two distinct groups of TTG suites (low-
3539 Al and high-Al) derived from distinct mafic sources with or without residual garnet. The low-
3540 Al gneisses plot in the field defined for the melts derived from depleted mafic source at shallow
3541 level without any significant residual garnet. Arrows in high-Al TTG field indicate batch of
3542 melting of mafic source at deeper levels transforming into garnet bearing amphibolites residue
3543 with 7–30% garnet in residue.
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Table captions

[Table 1](#) Summary table of interpreted zircon U-Pb age and Lu-Hf characteristics for studied granitoids.

Supplementary Table captions

[Supplementary Table 1](#). Major oxide (wt.%), trace and REE (ppm) concentrations of Holenarsipur greenstone belt and other greenstone belts in the studies cratonic core region.

[Supplementary Table 2](#). Major oxide (wt.%), trace and REE (ppm) concentrations of TTG-type granitoids surrounding HGB and other western TTGs in the studies cratonic core region.

[Supplementary Table 3](#). Zircon U-Pb data by LA-ICP-MS for the TTGs surrounding Holenarsipur greenstone belt along with reference material analysis.

[Supplementary Table 4](#). *in-situ* Lu-Hf isotope data for the TTGs surrounding Holenarsipur greenstone belt along with those for reference material.

[Supplementary Table 5](#). Sm-Nd isotopic data for rocks of the southwestern, northcentral and eastern blocks of the Holenarsipur greenstone belt.

[Supplementary Table 6](#). Operating conditions for U-Pb and Lu-Hf isotope measurements using LA-ICP-MS and LA-MC-ICP-MS.

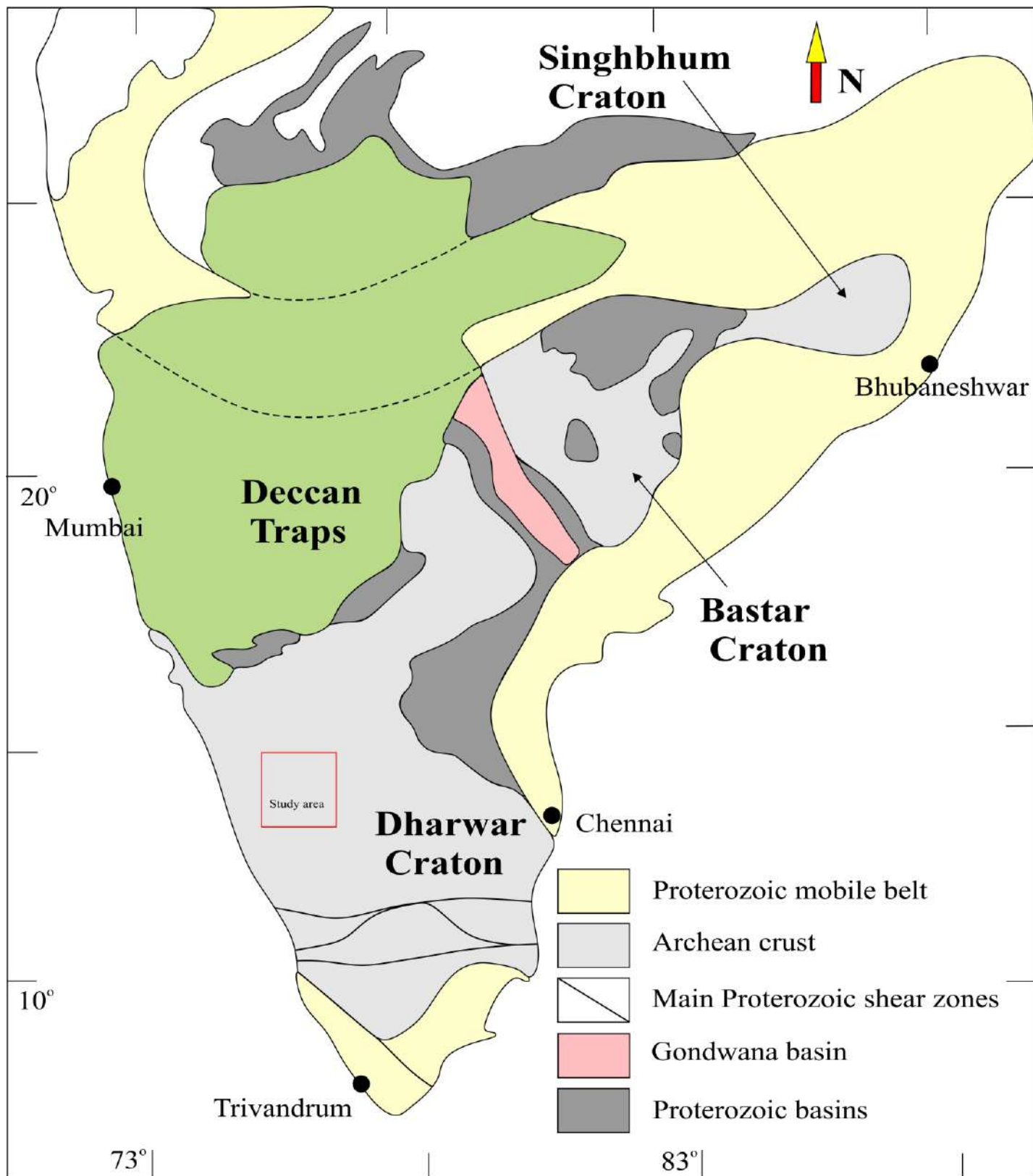


Fig. 1

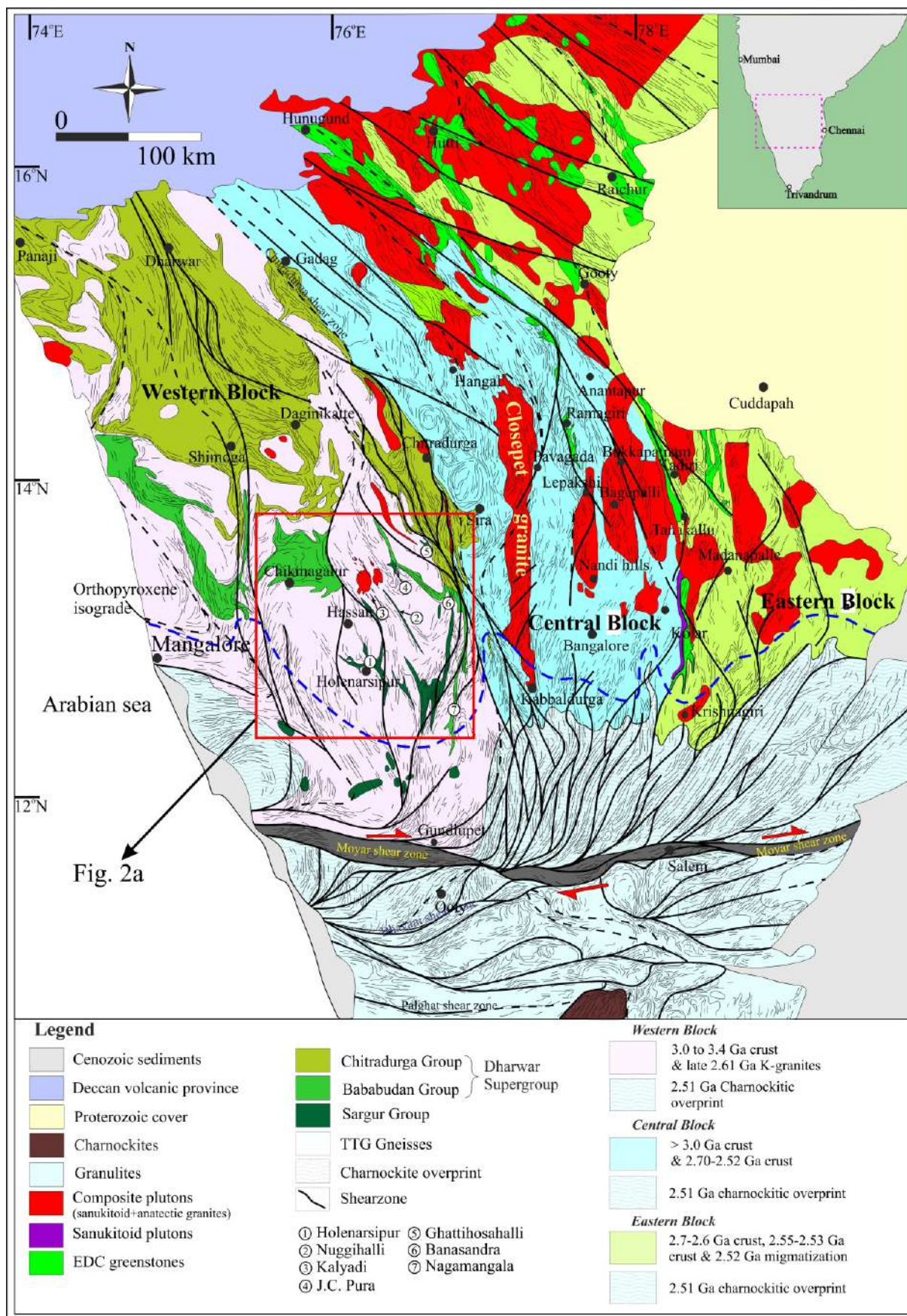


Fig. 2

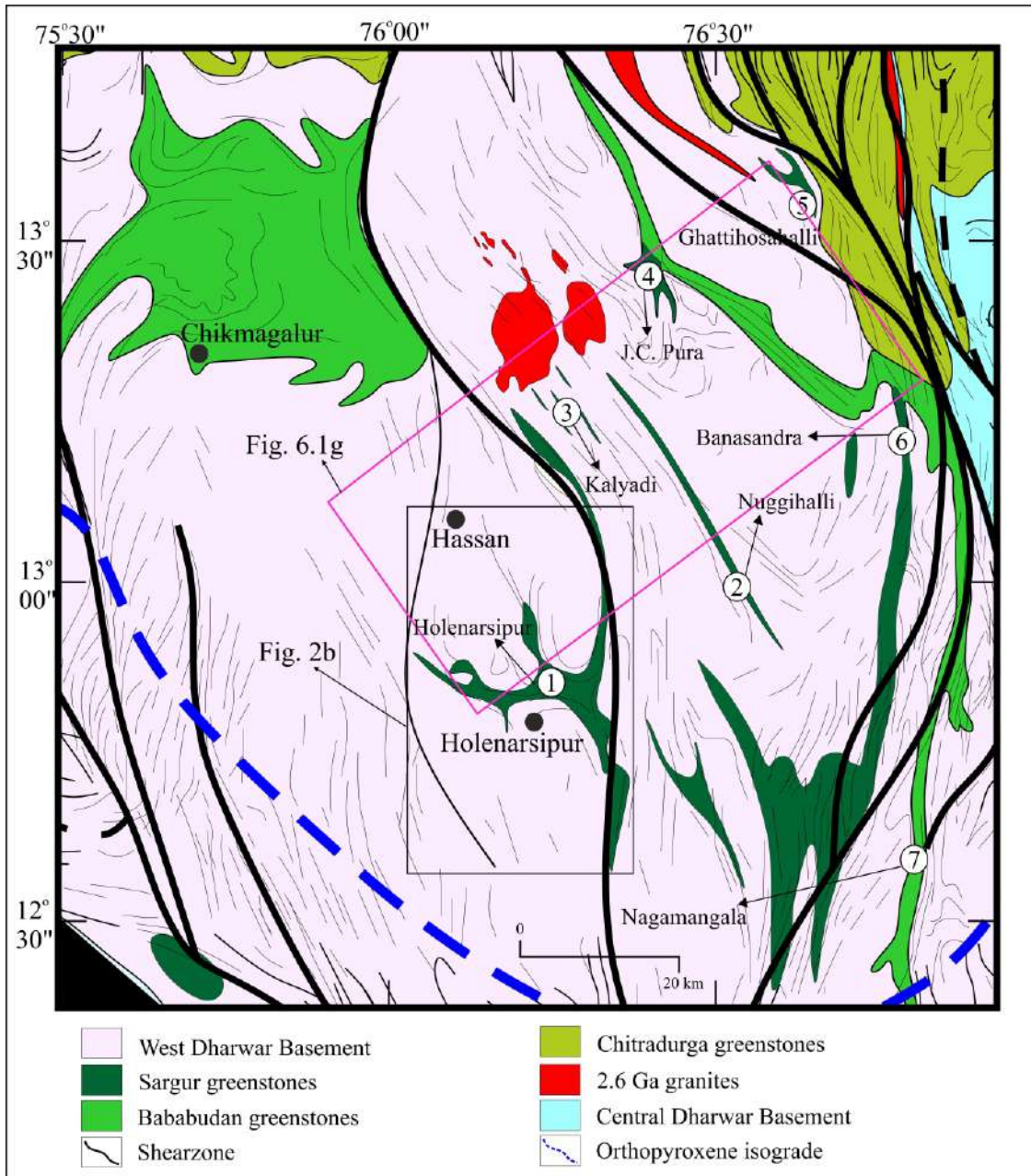


Fig. 2a

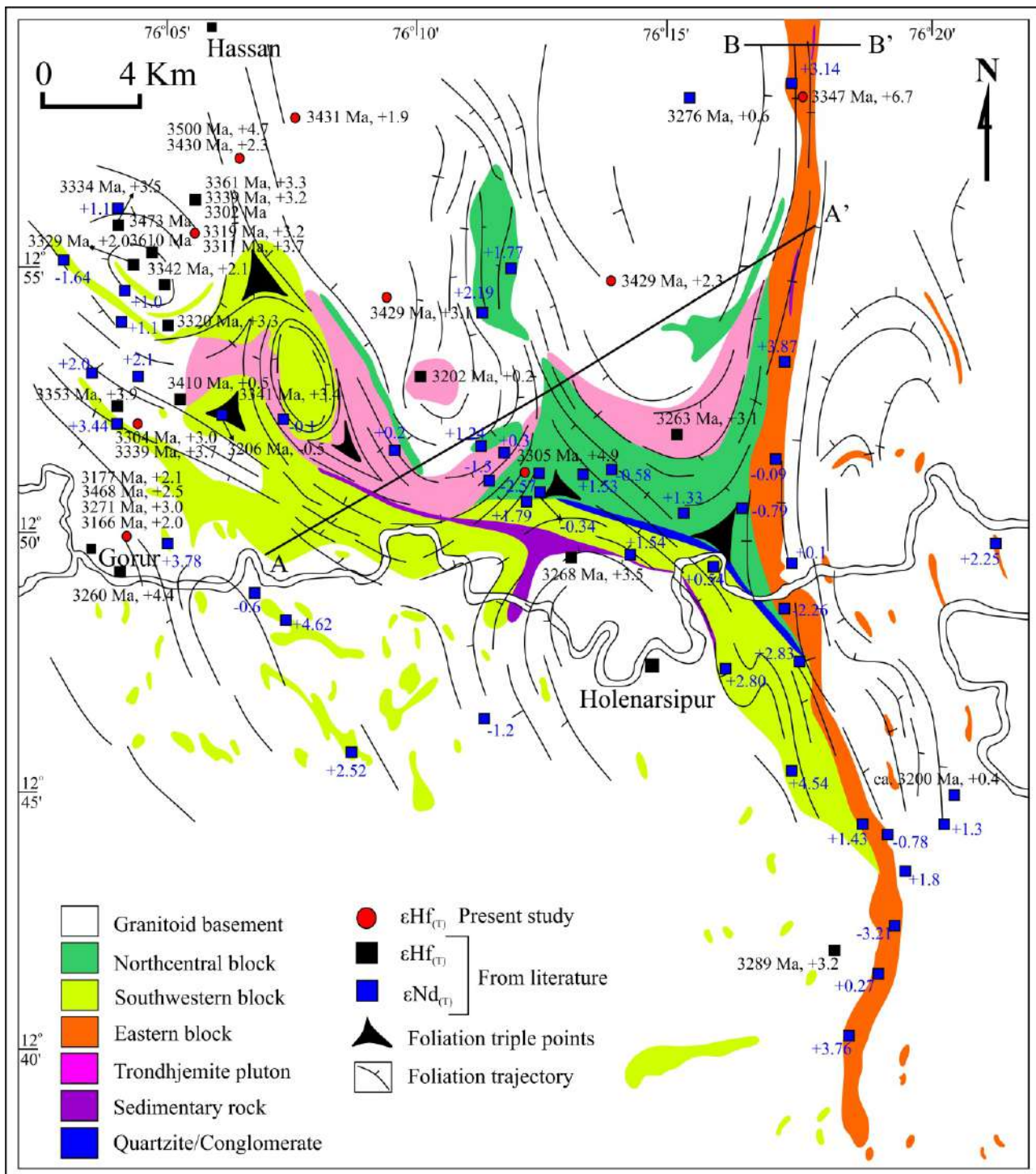


Fig. 2b

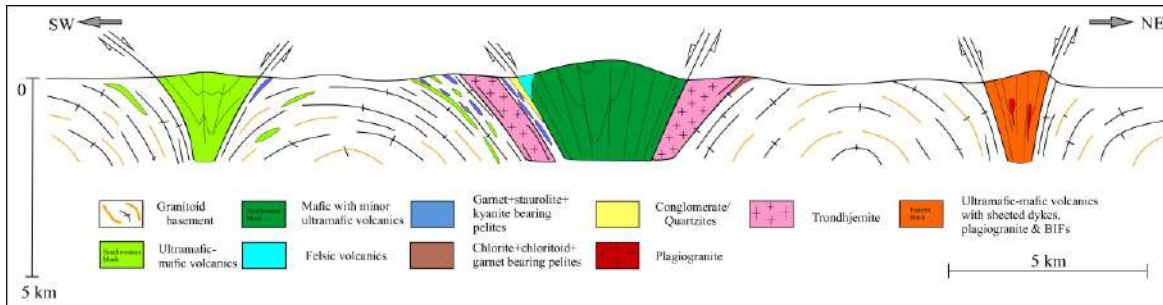


Fig. 2c



Fig. 3.1a-b

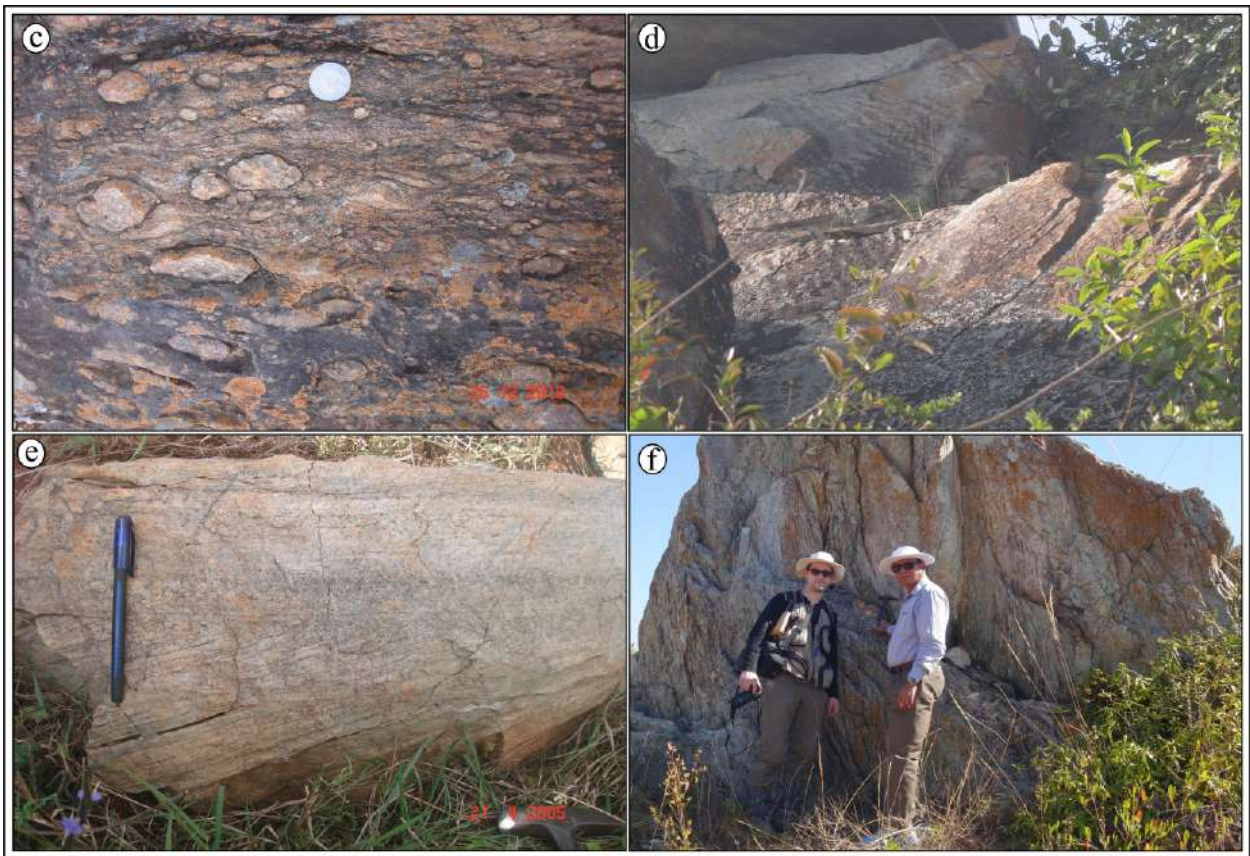


Fig. 3.1c-f

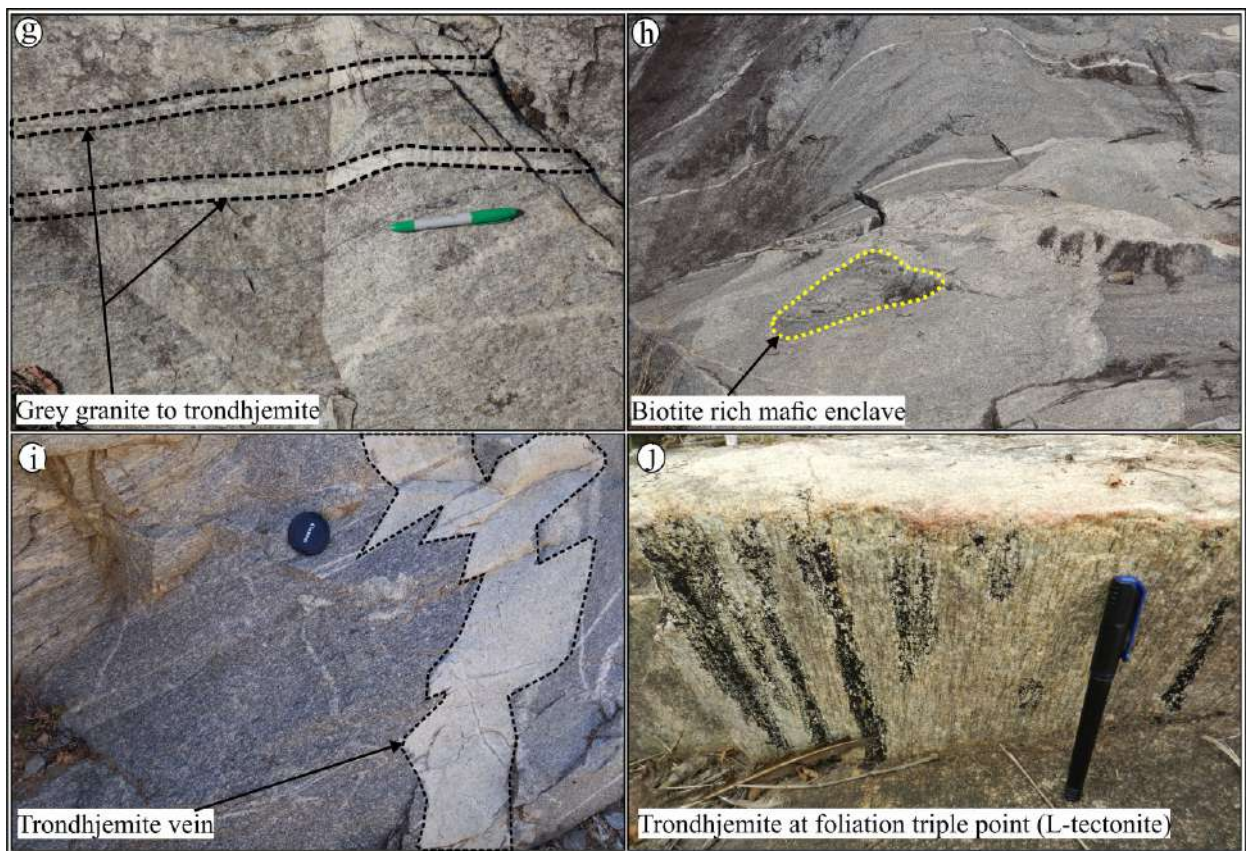


Fig. 3.1g-j

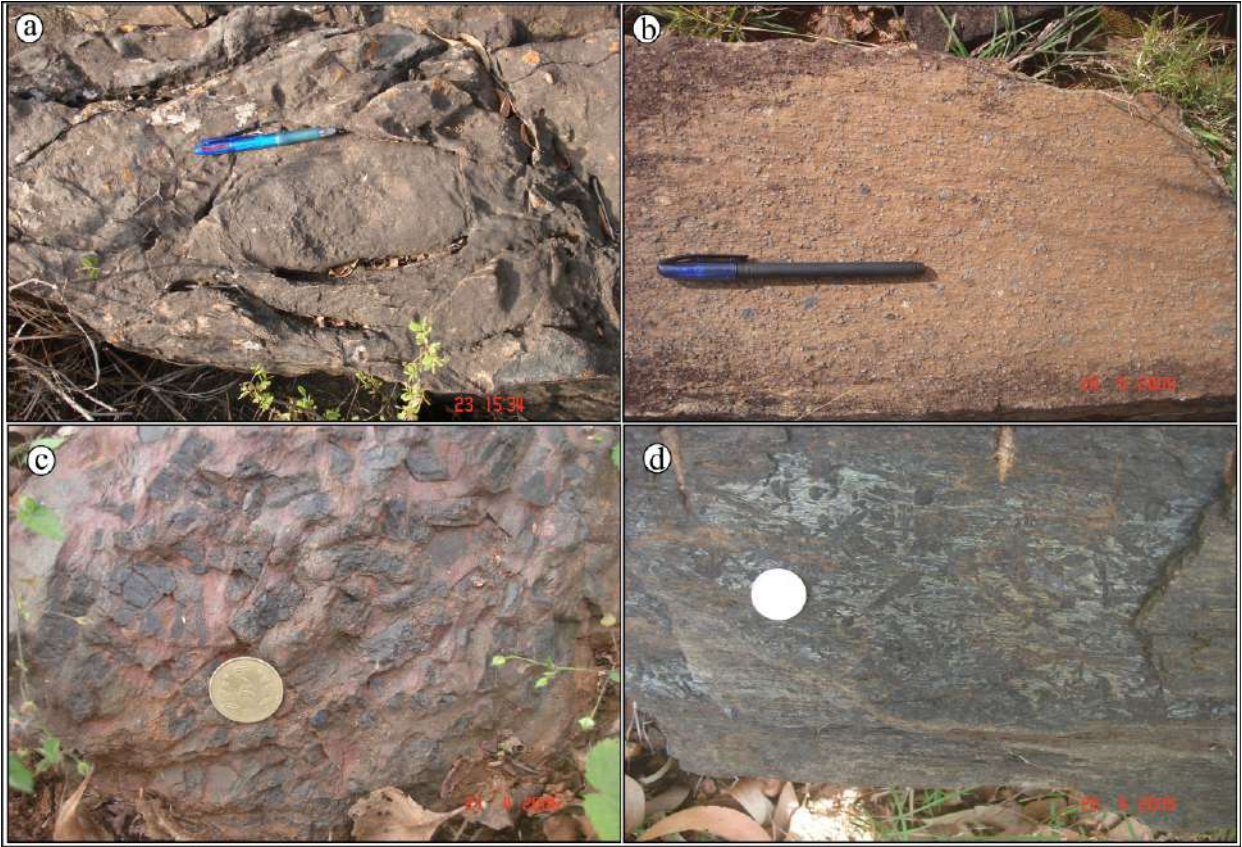


Fig. 3.2a-d

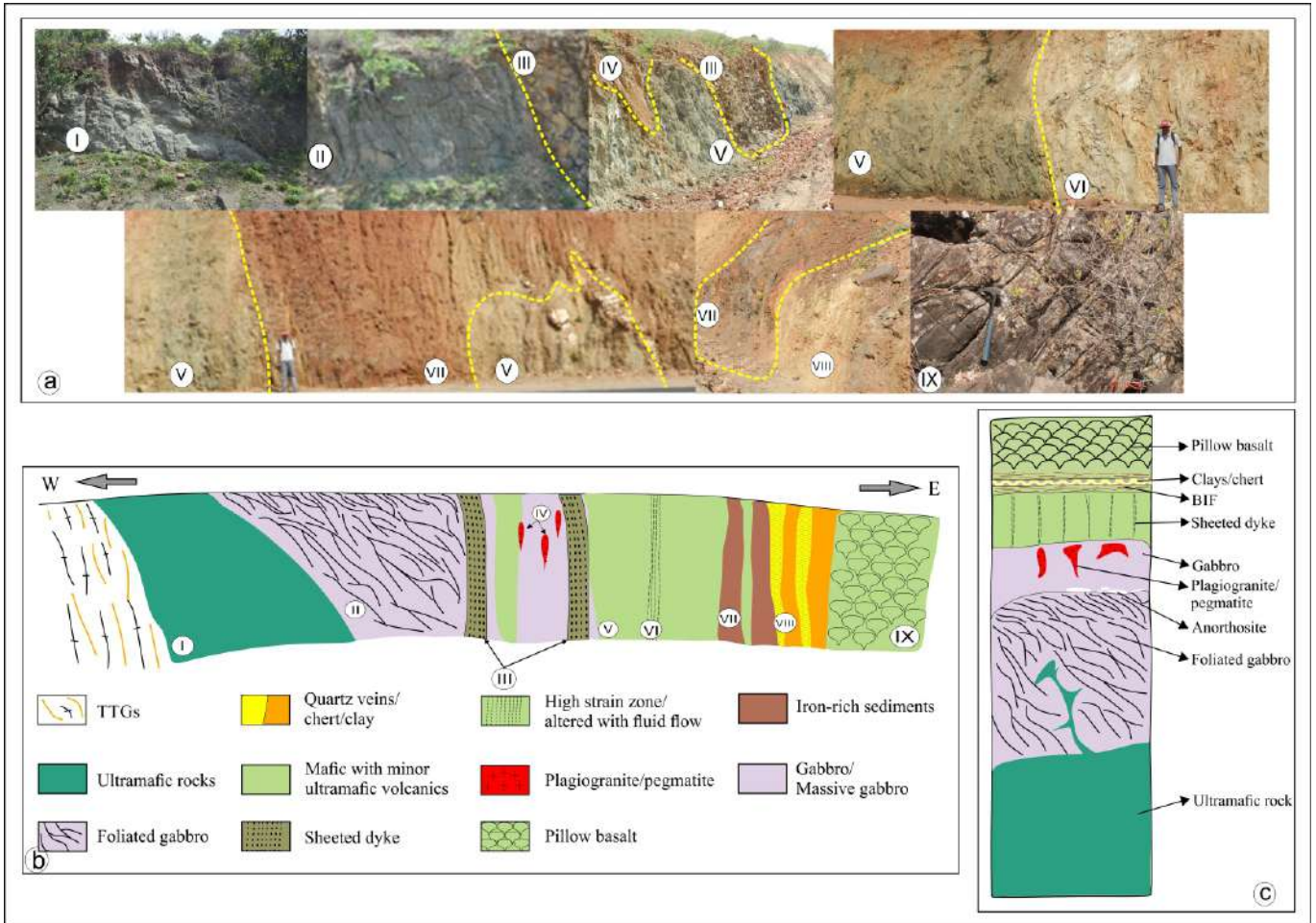


Fig. 3.3a-c

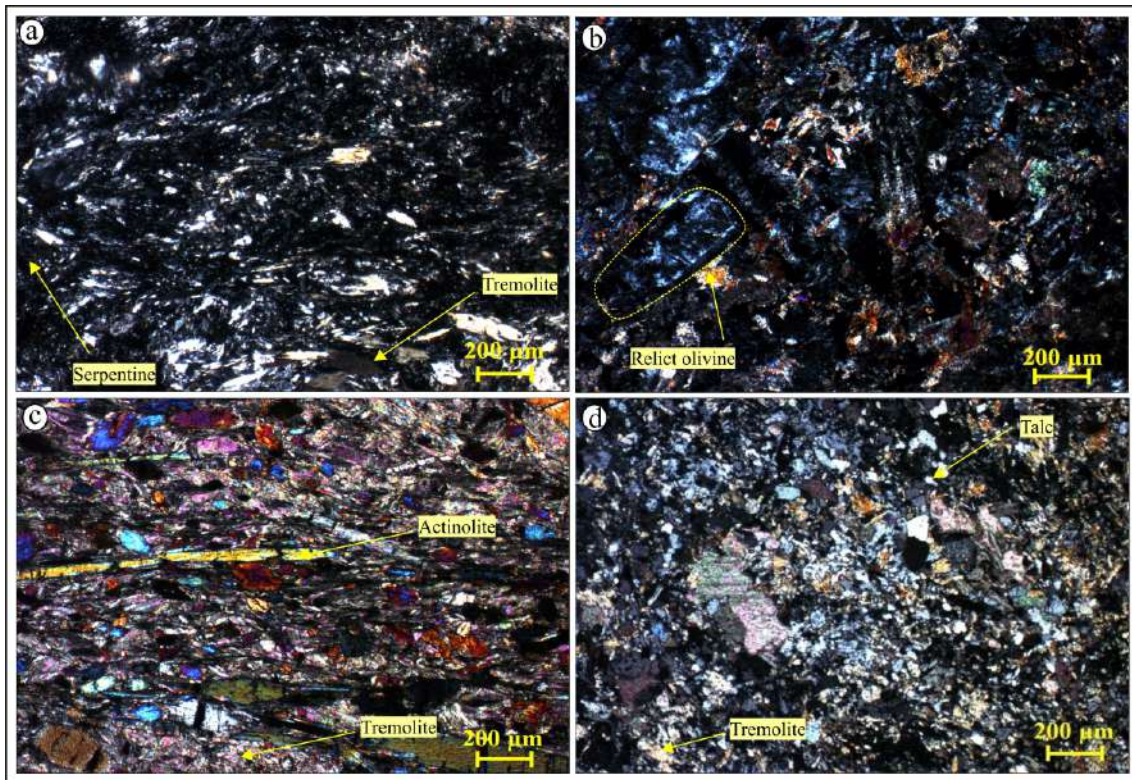


Fig. 4.1a-f

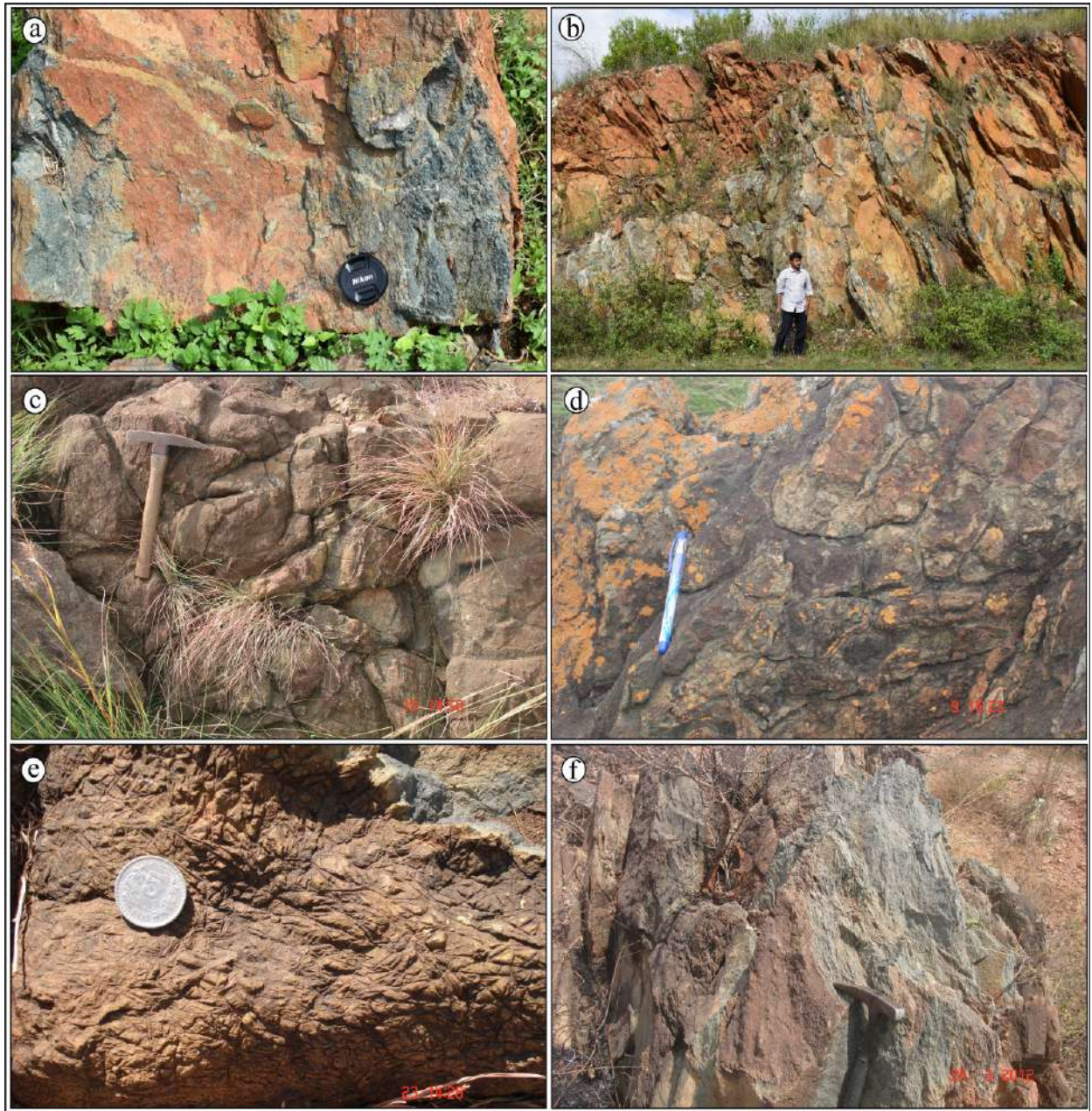


Fig. 5.1a-f

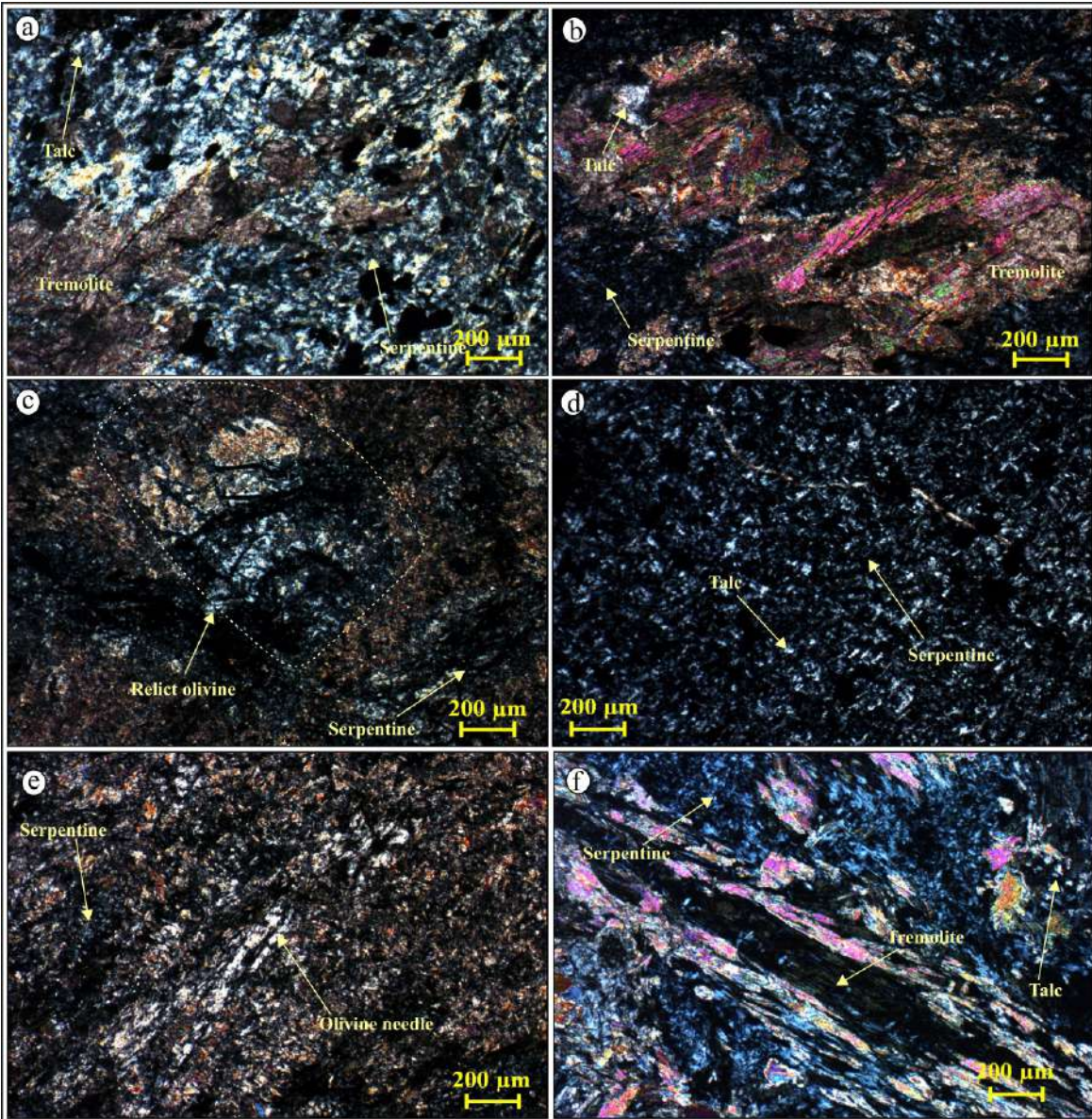


Fig. 5.2 a-f

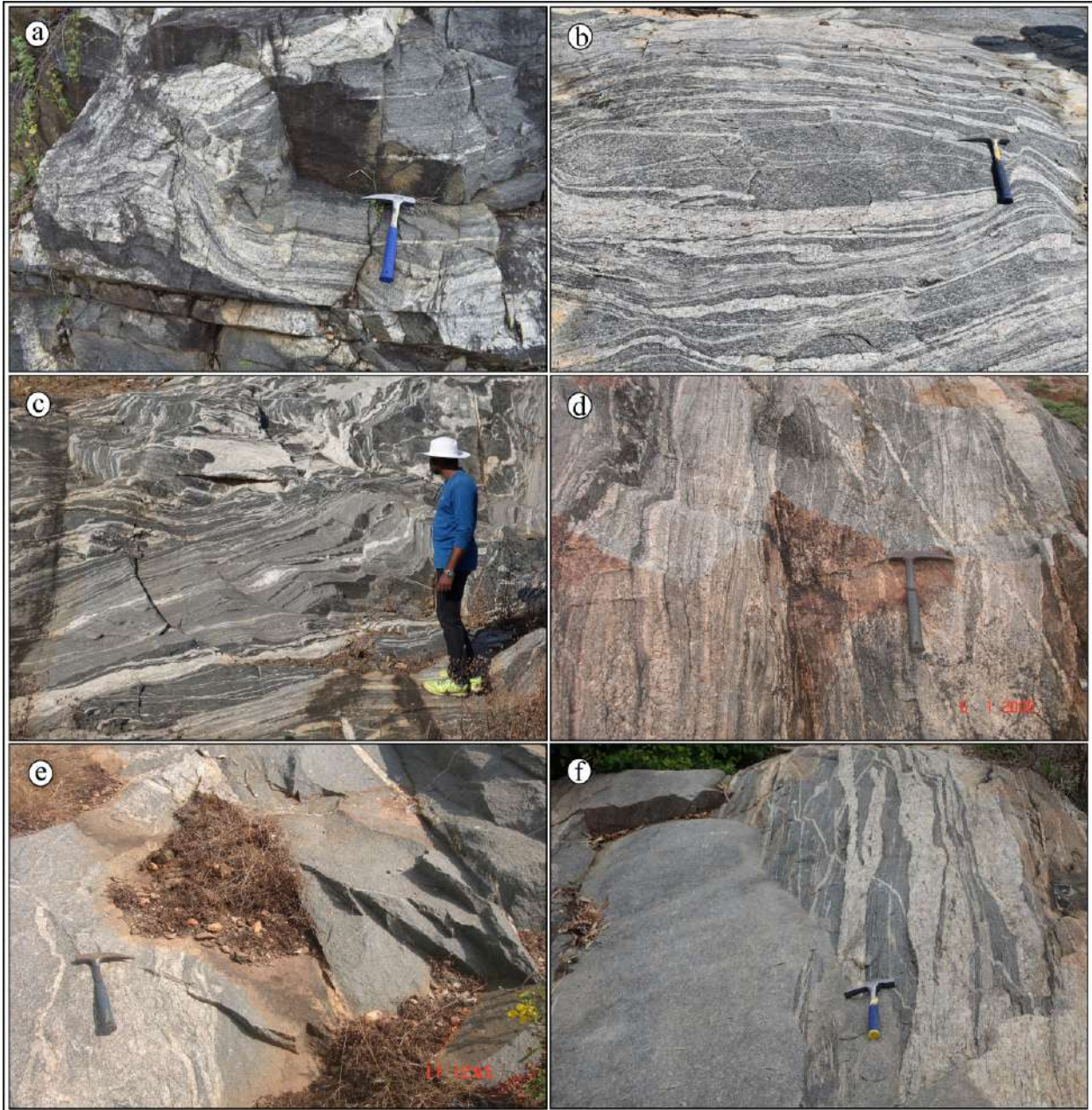


Fig. 6.1 a-f

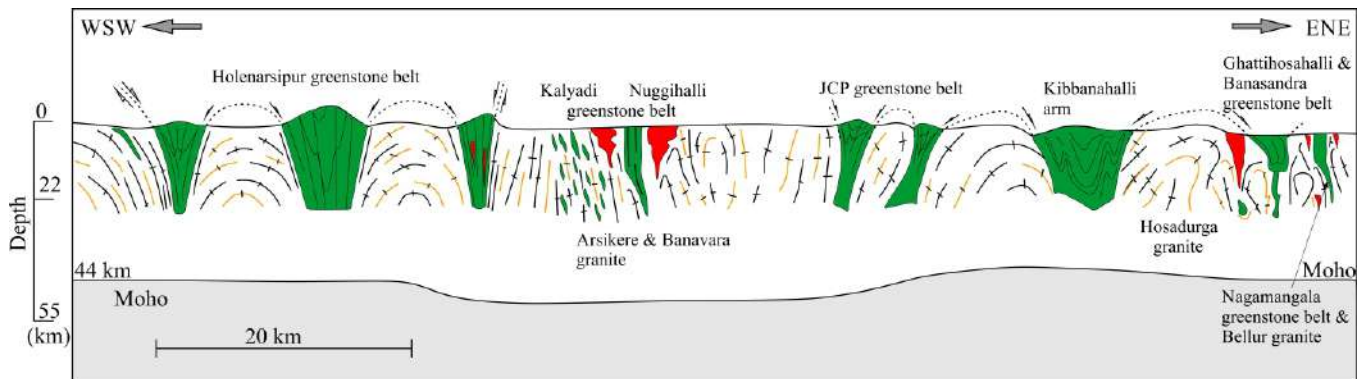


Fig. 6.1 g

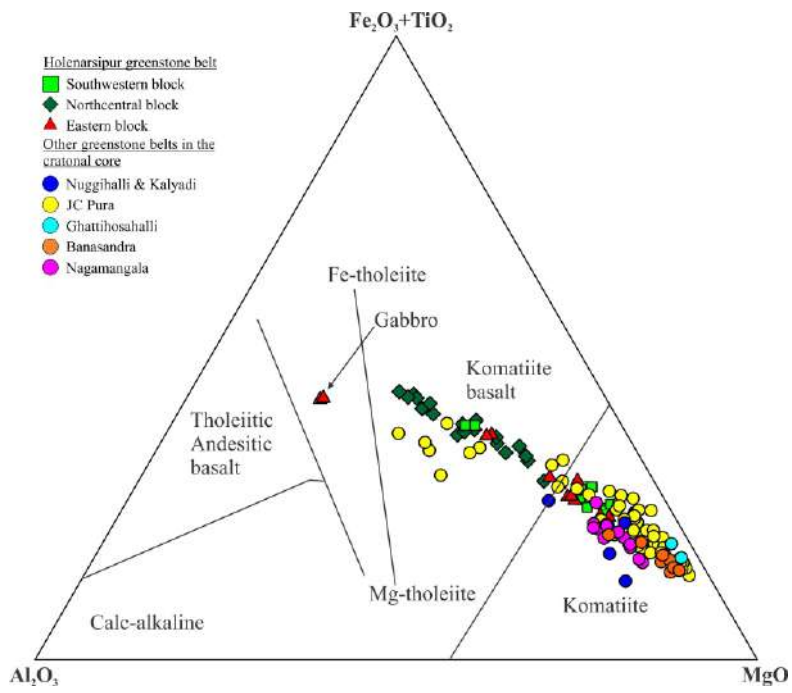


Fig. 7.1a

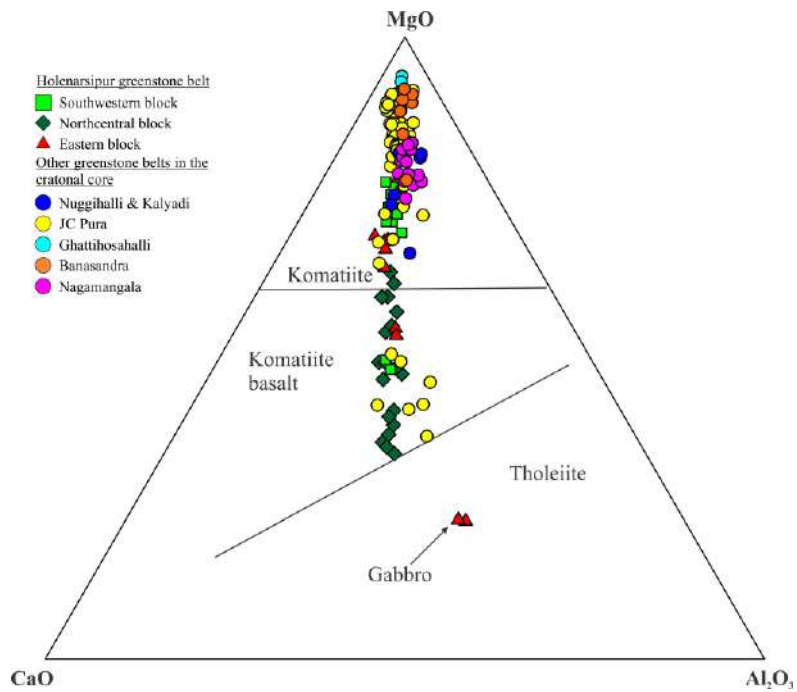


Fig. 7.1b

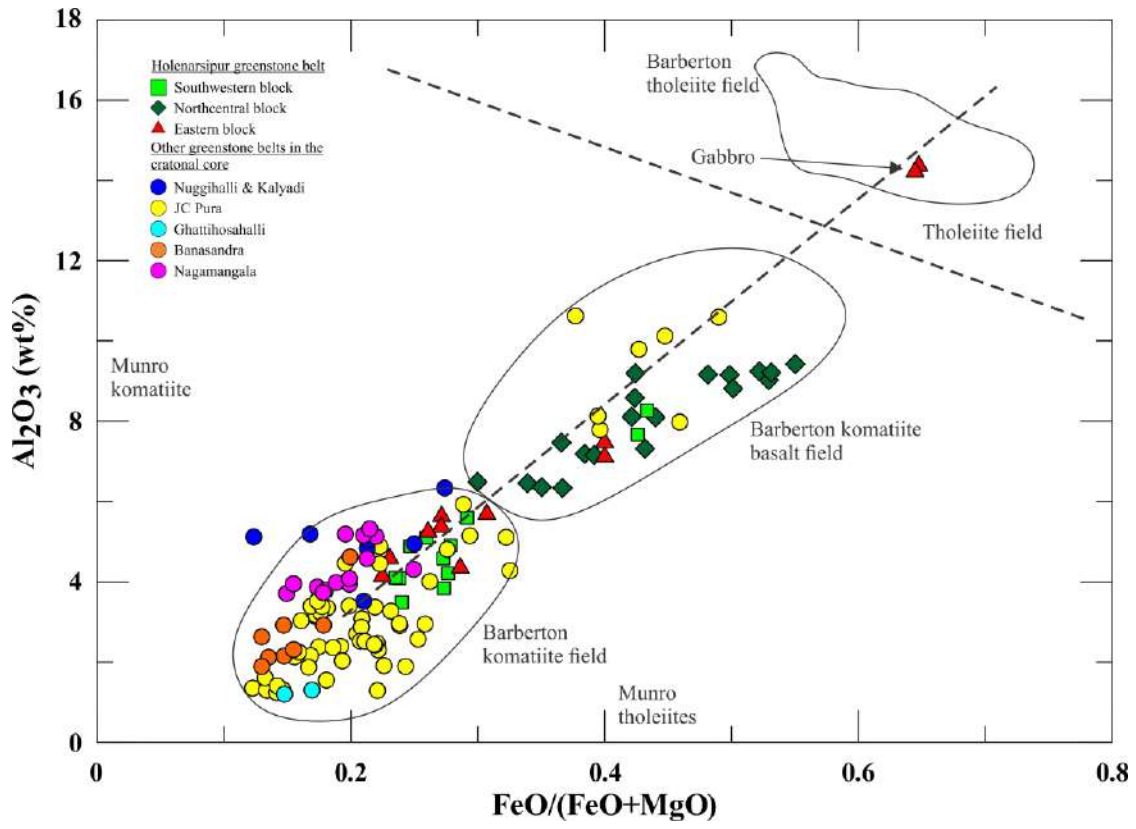


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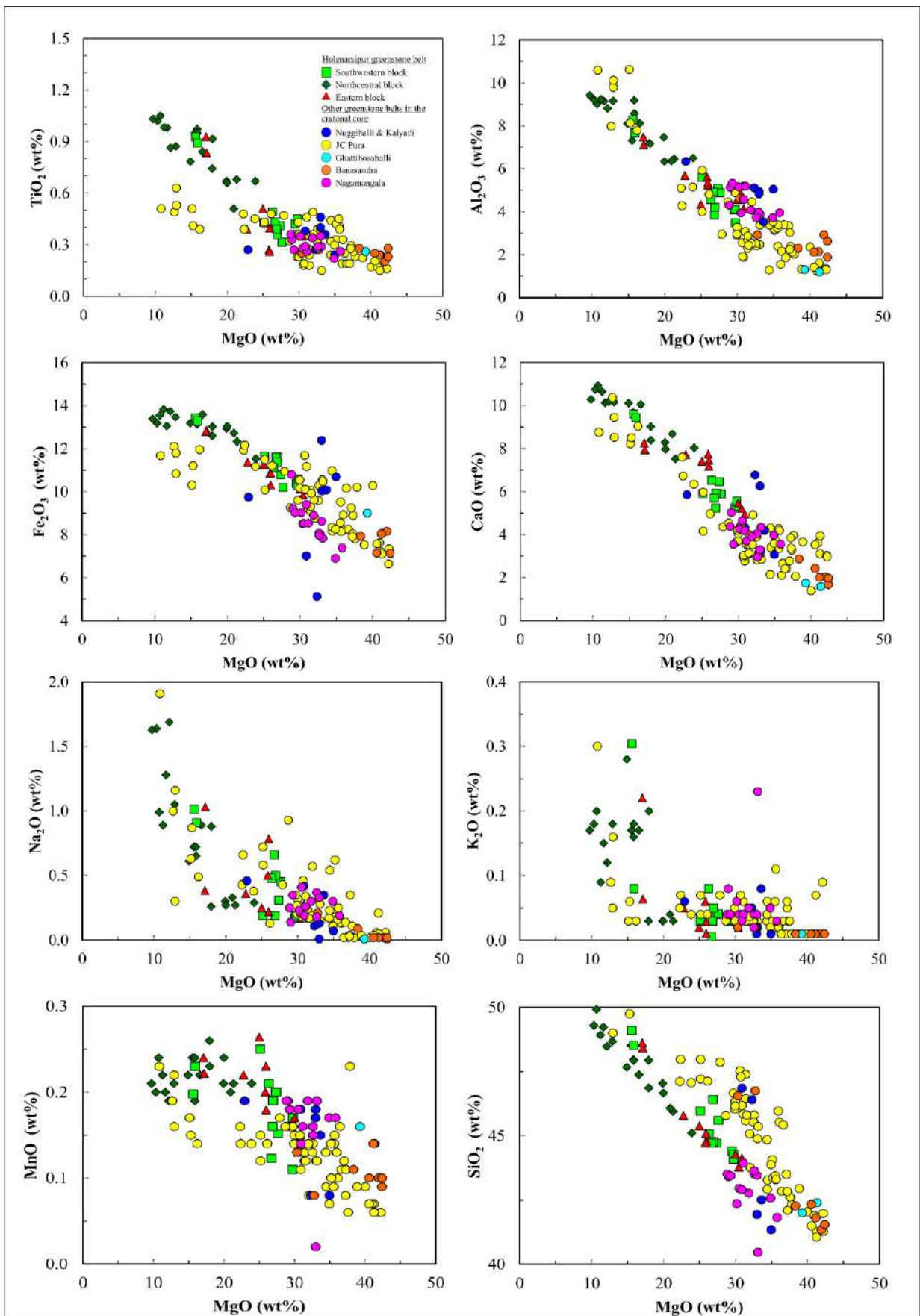


Fig. 7.1d

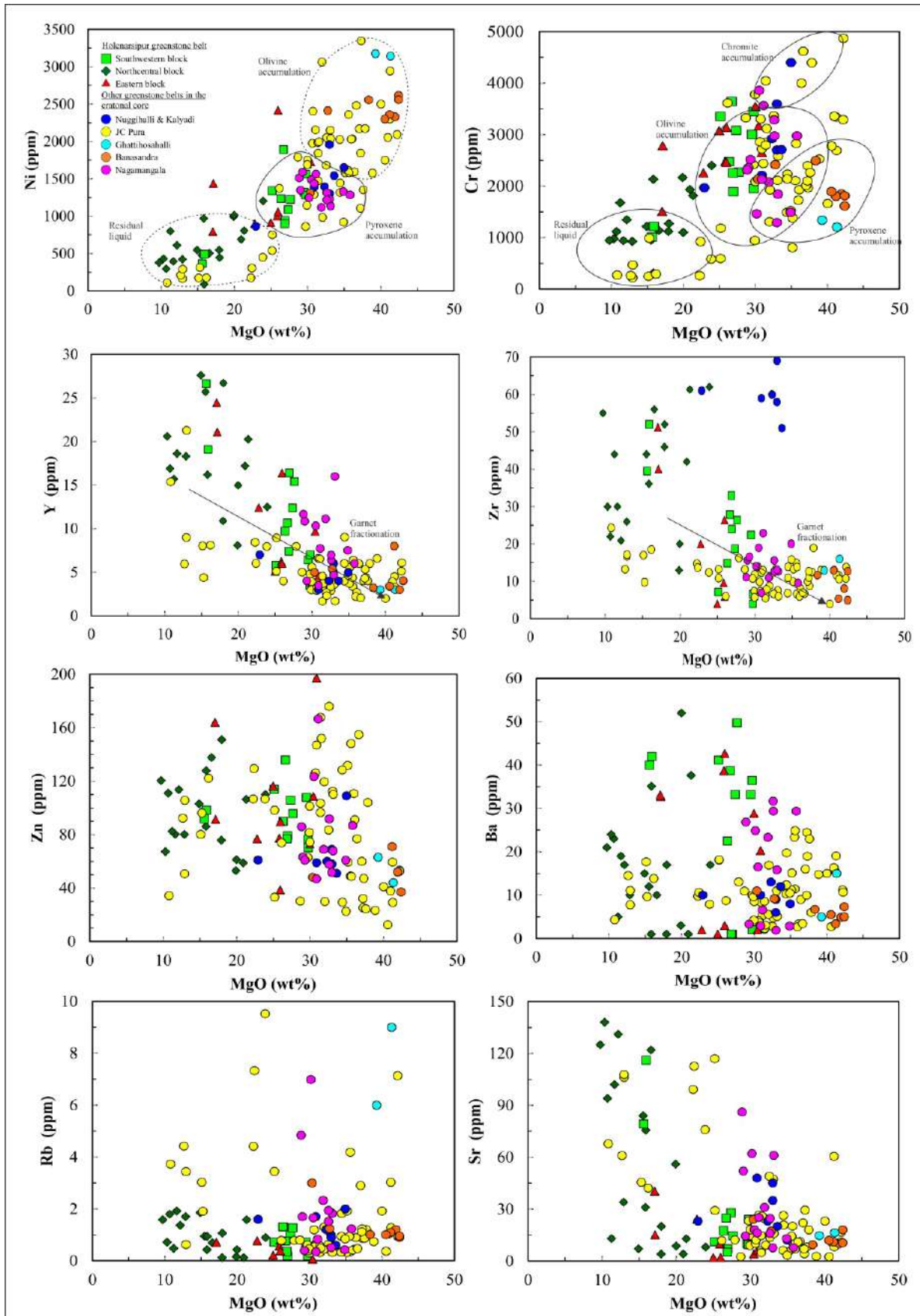


Fig. 7.1e

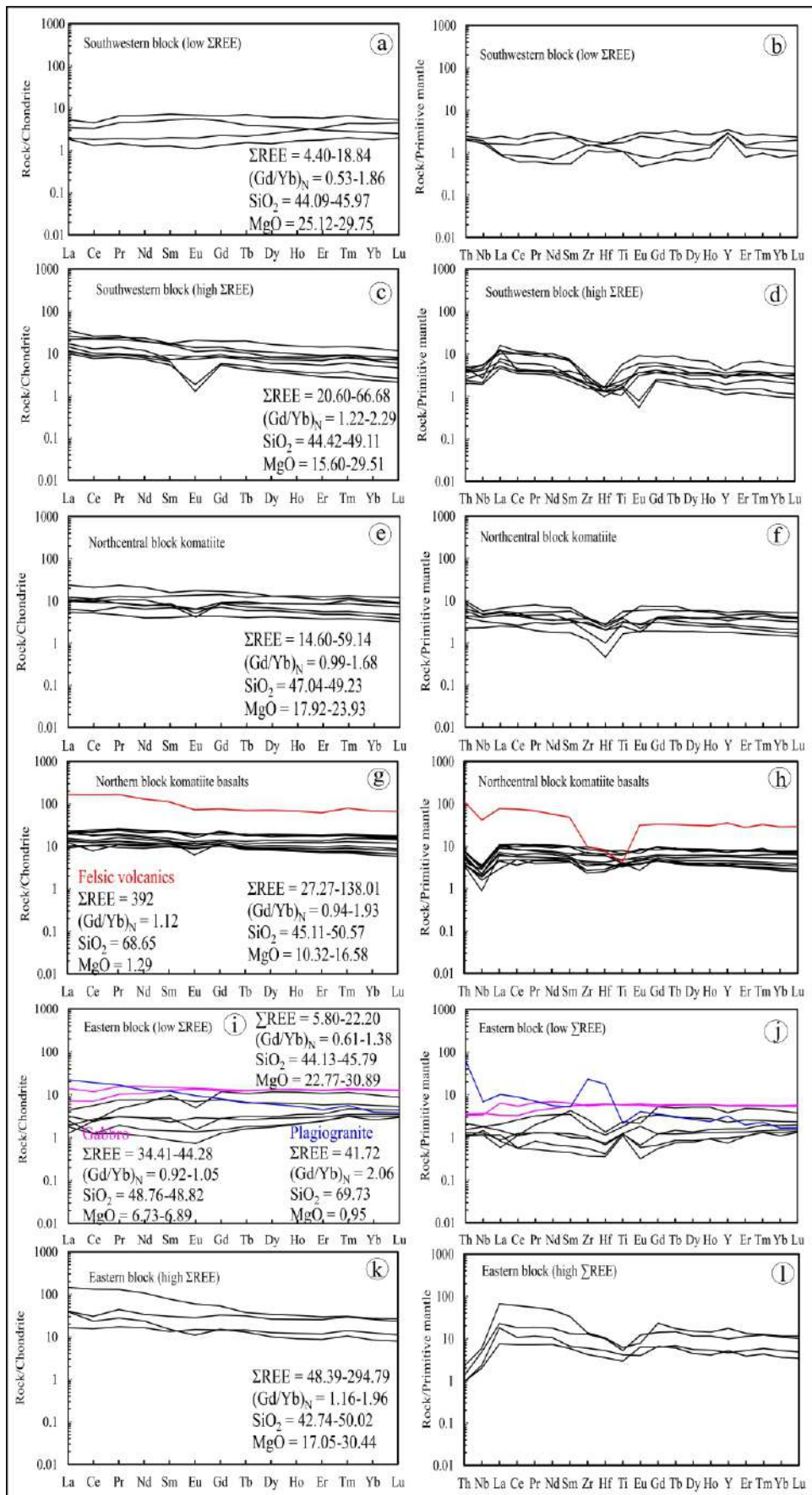


Fig. 7.2a-1

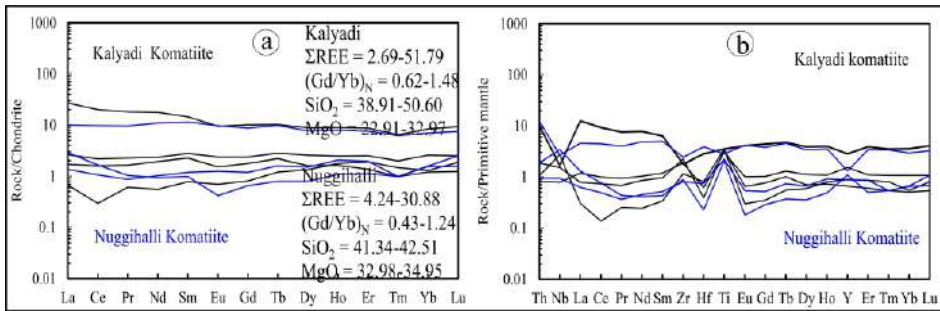


Fig. 7.3a-b

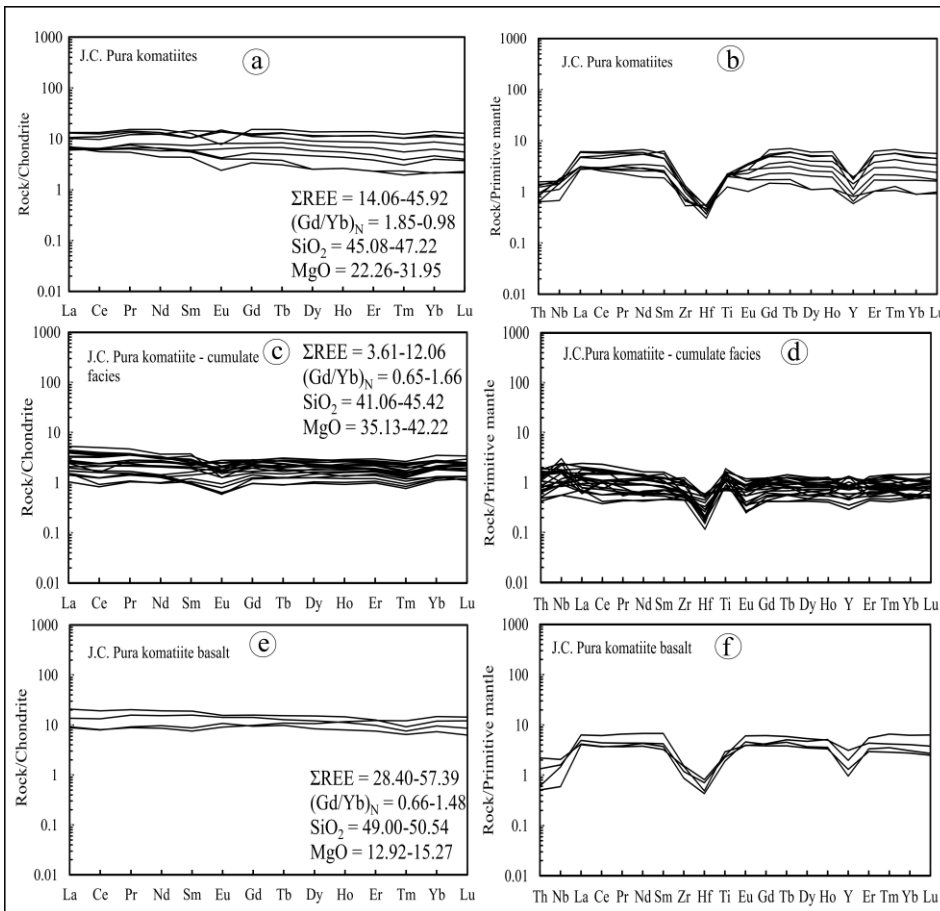


Fig. 7.4a-f

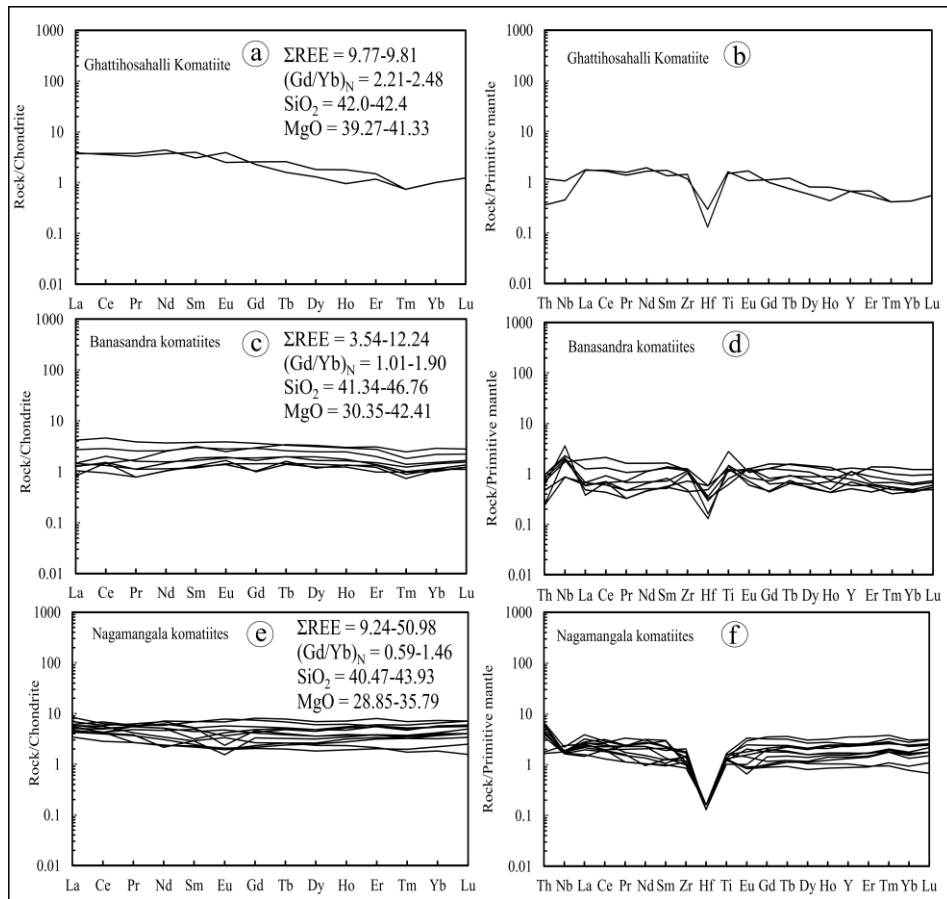


Fig. 7.5a-f

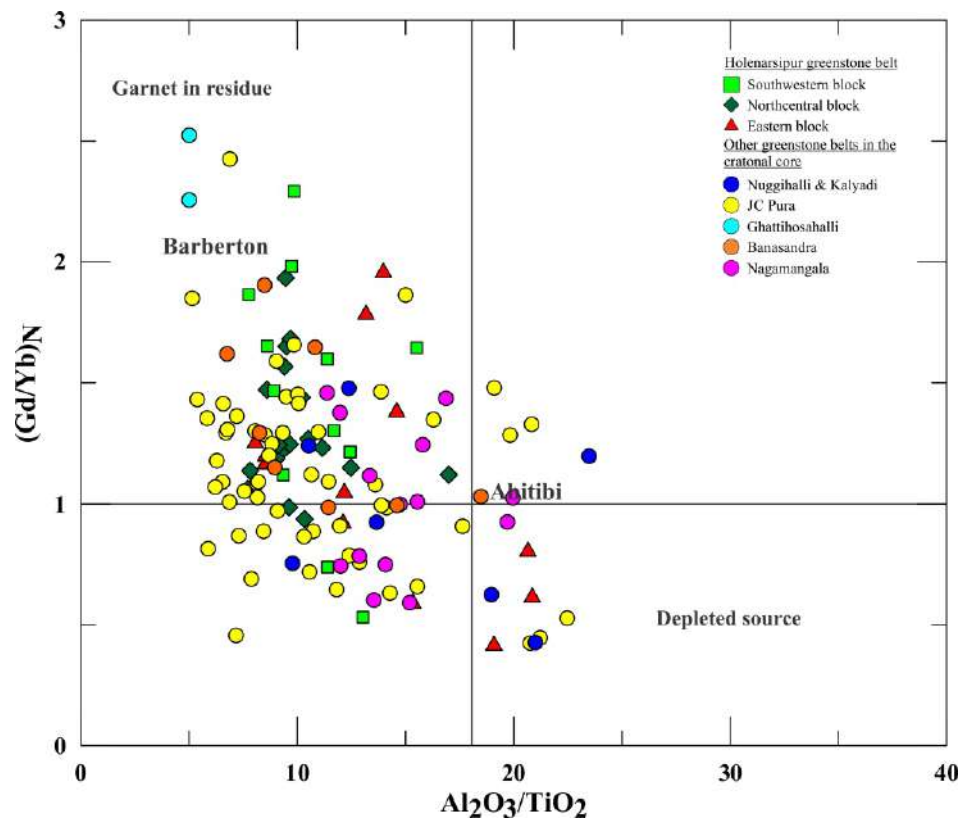


Fig. 7.6a

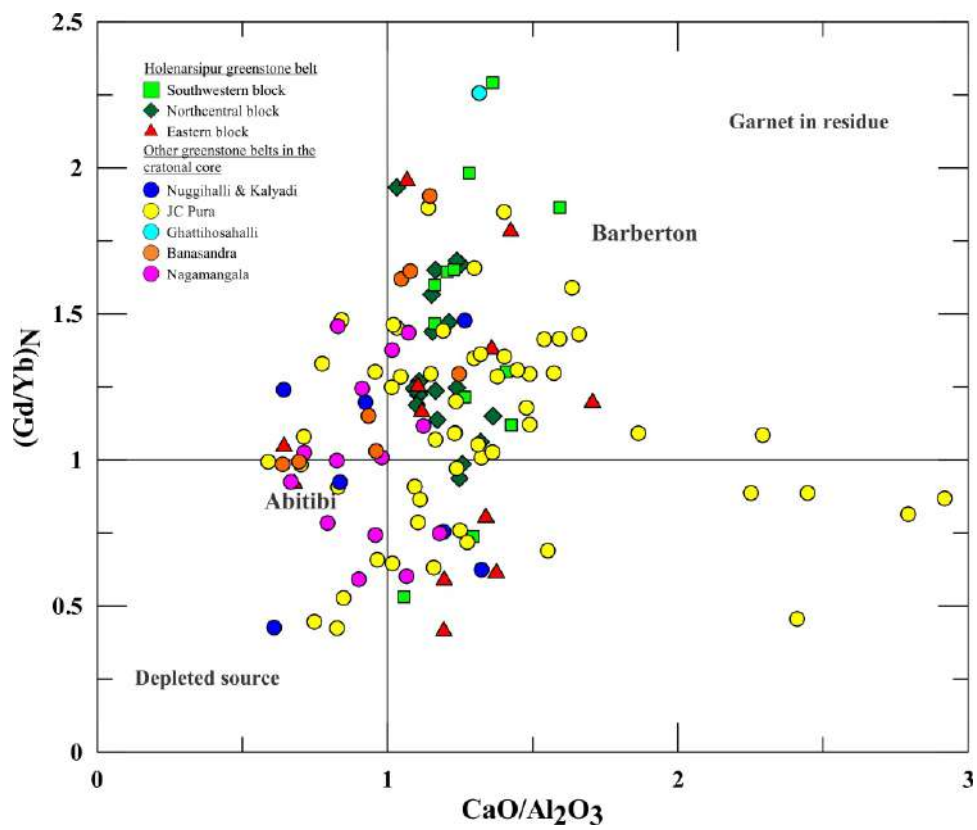


Fig. 7.6b

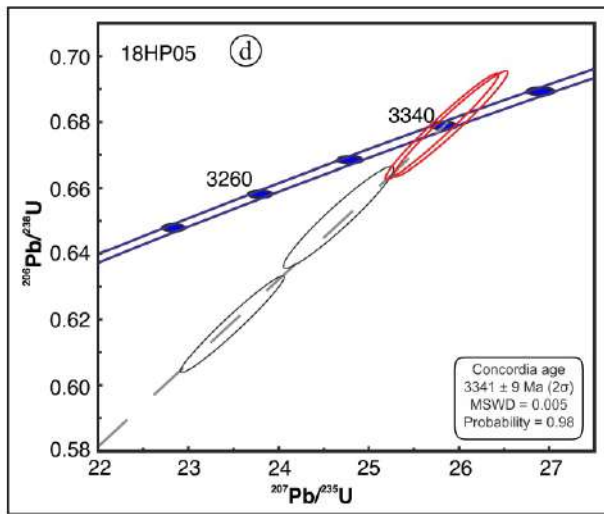
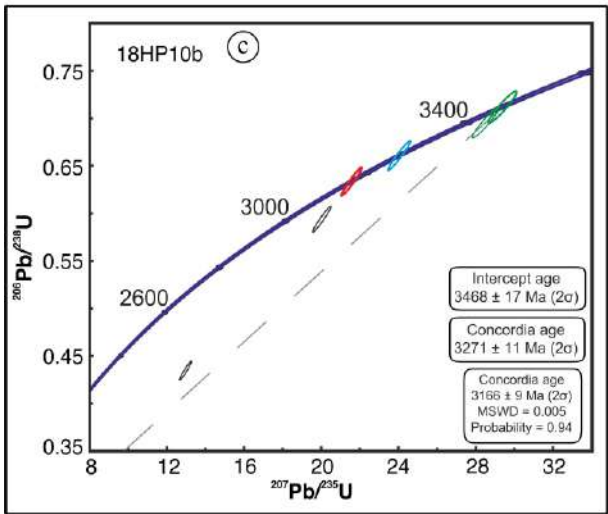
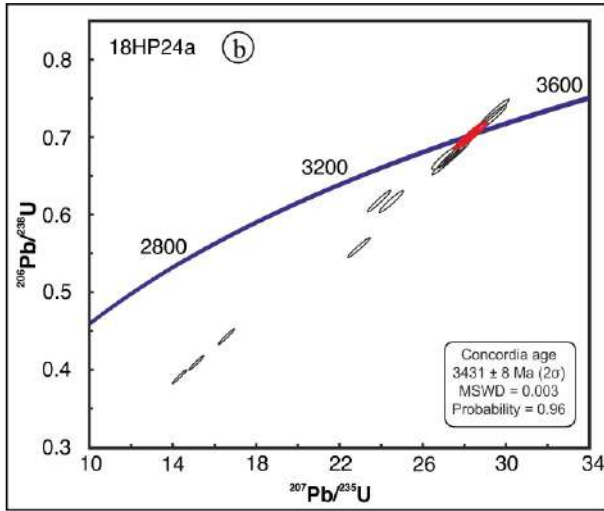
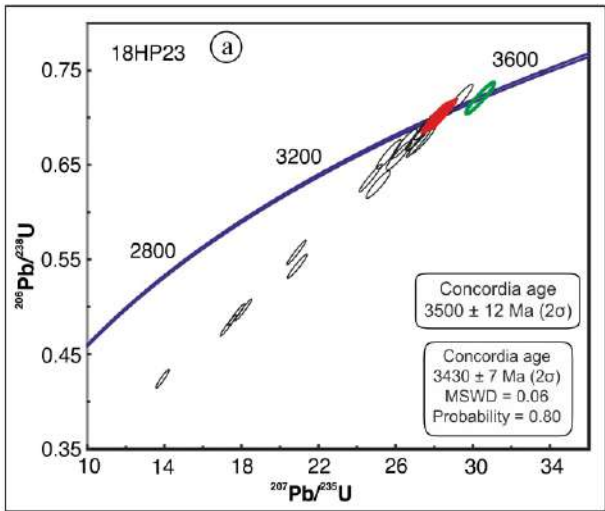


Fig. 7.7a-d

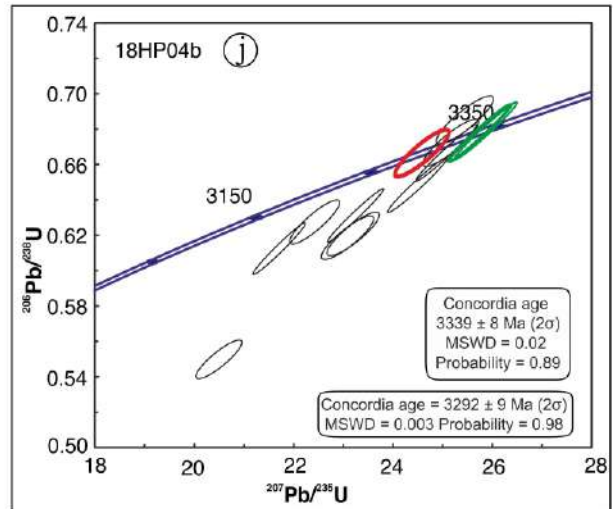
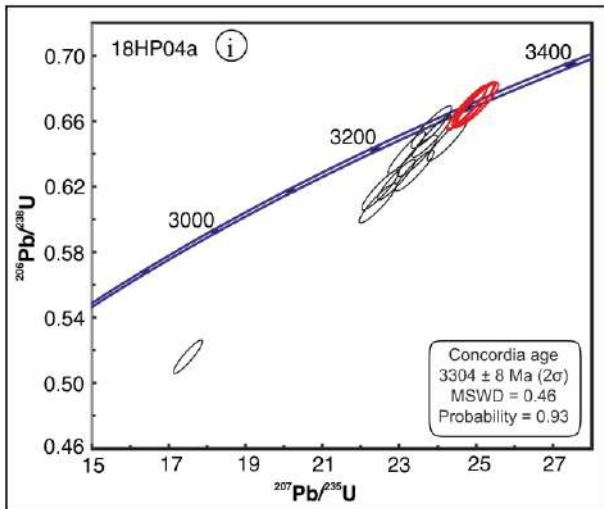
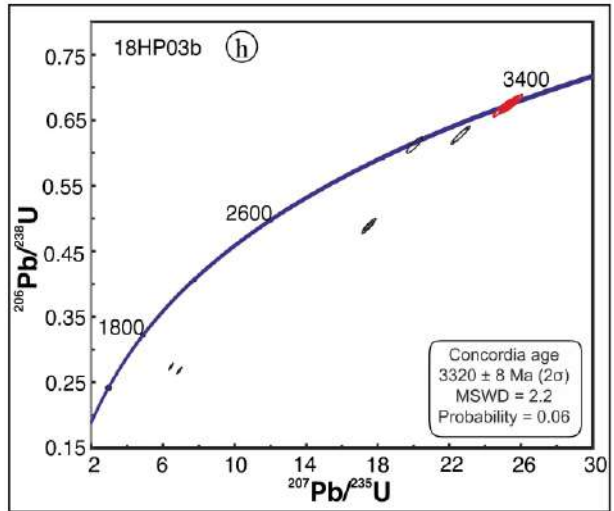
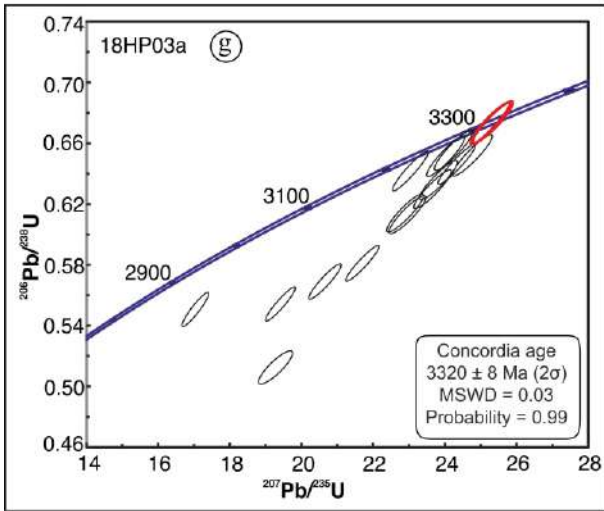
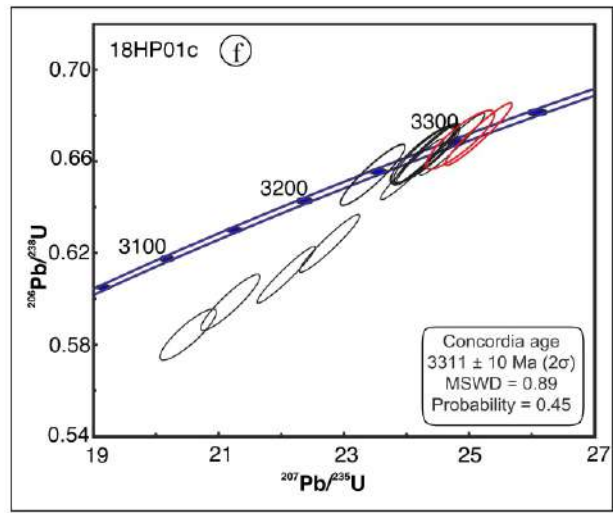
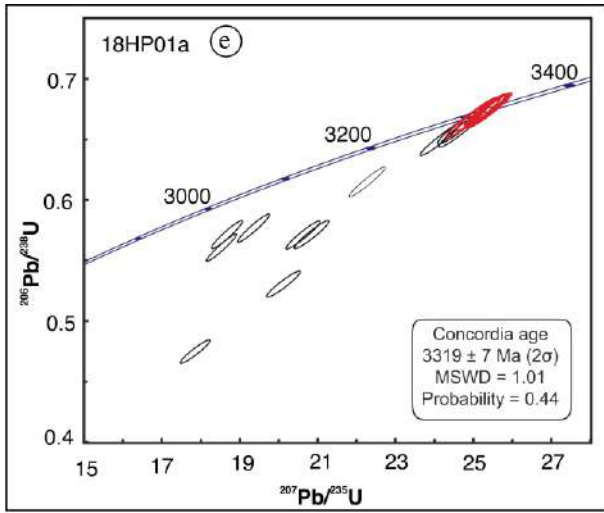


Fig. 7.7e-j

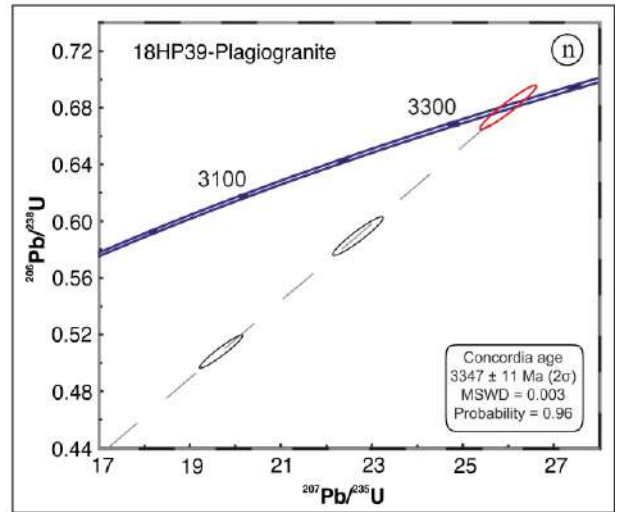
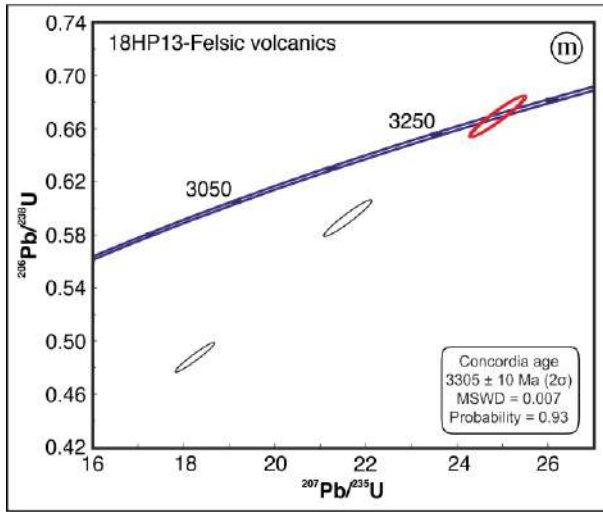
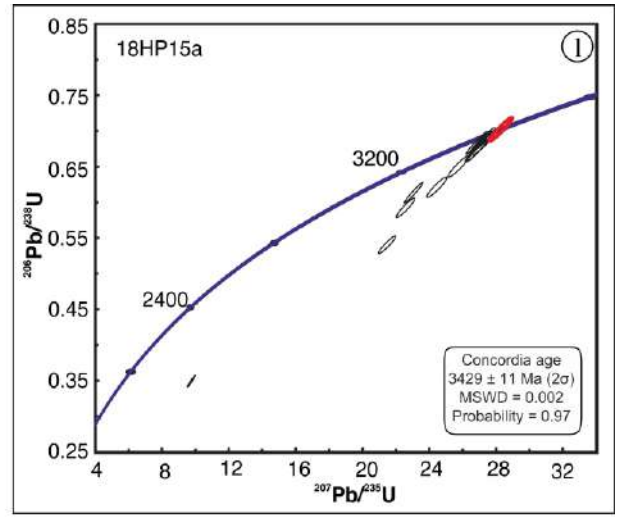
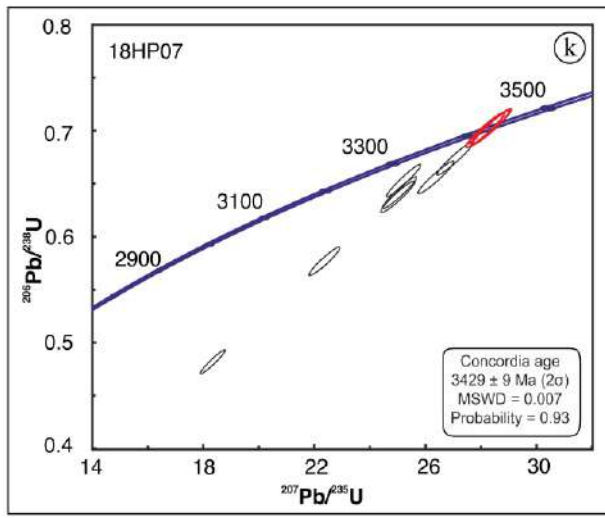


Fig. 7.7k-n

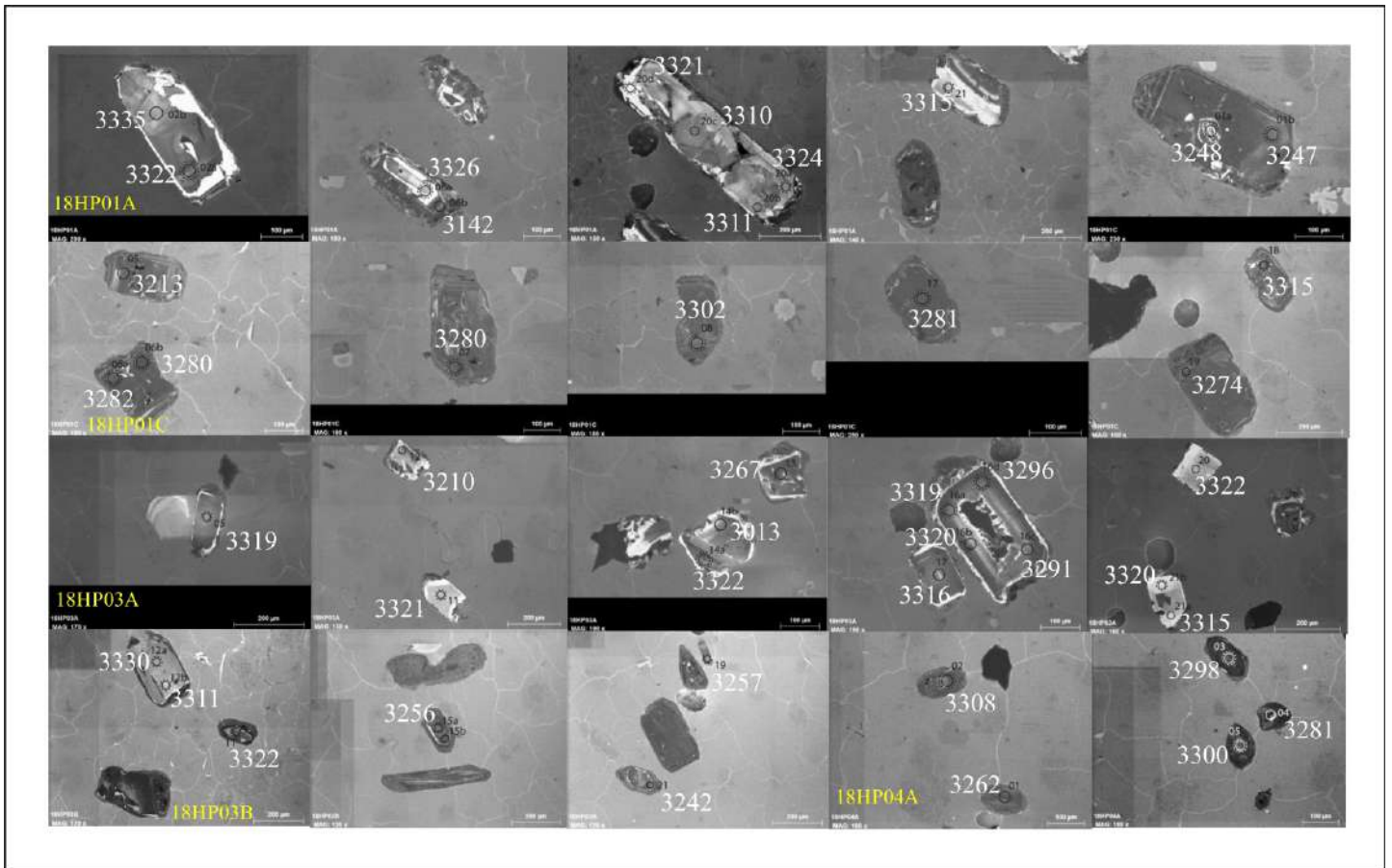


Fig. 7.8a

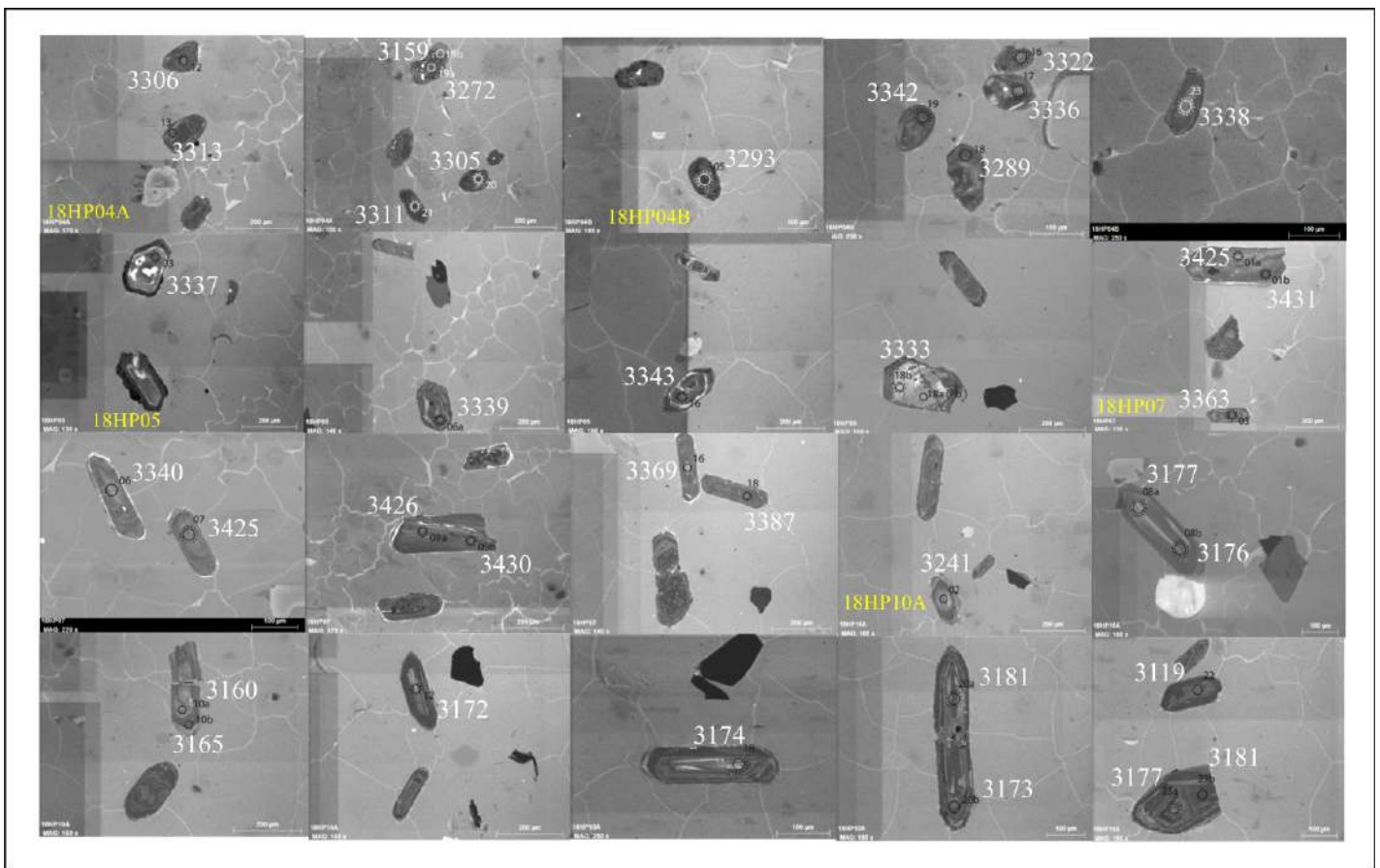


Fig. 7.8b

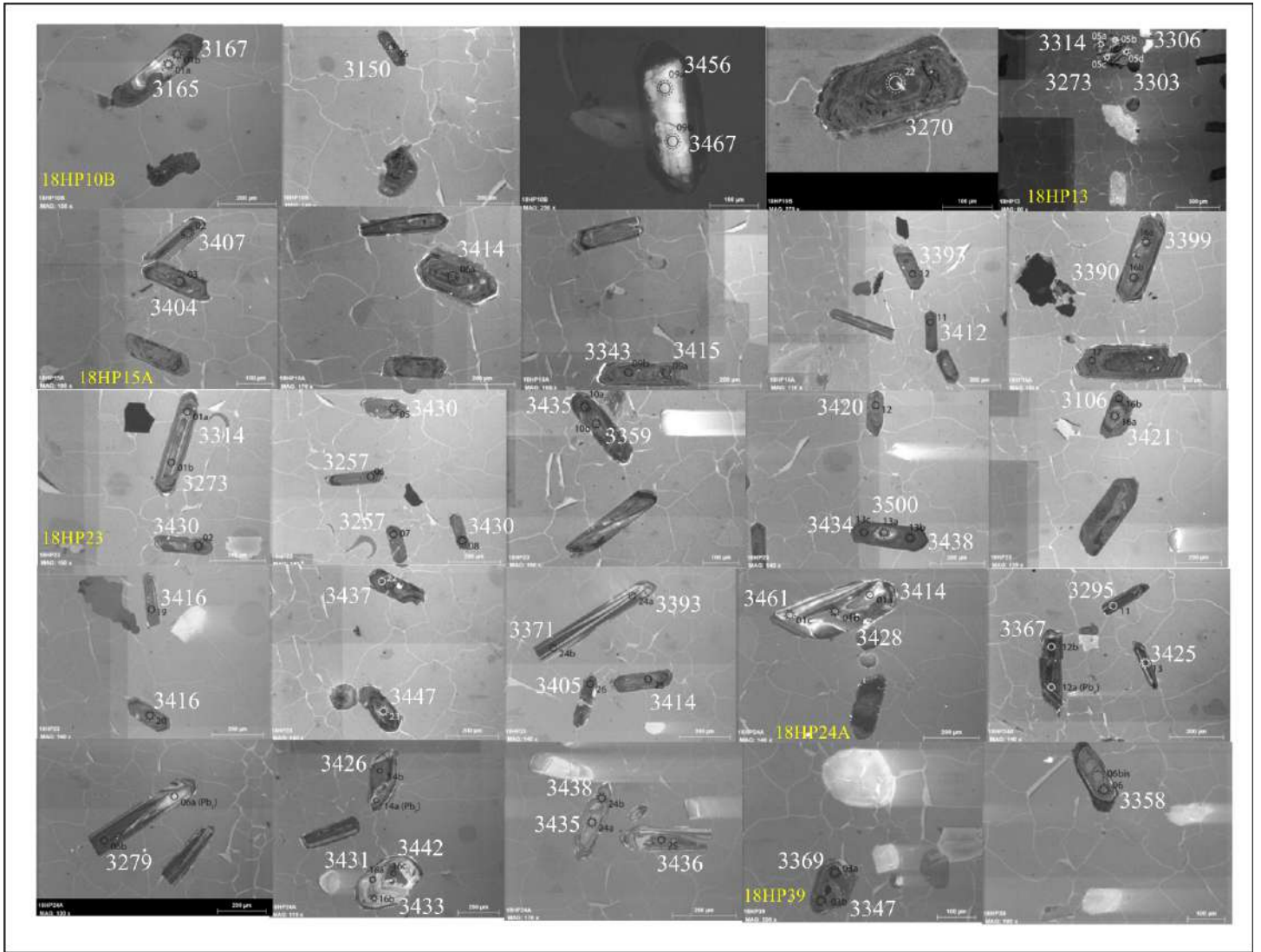


Fig. 7.8c

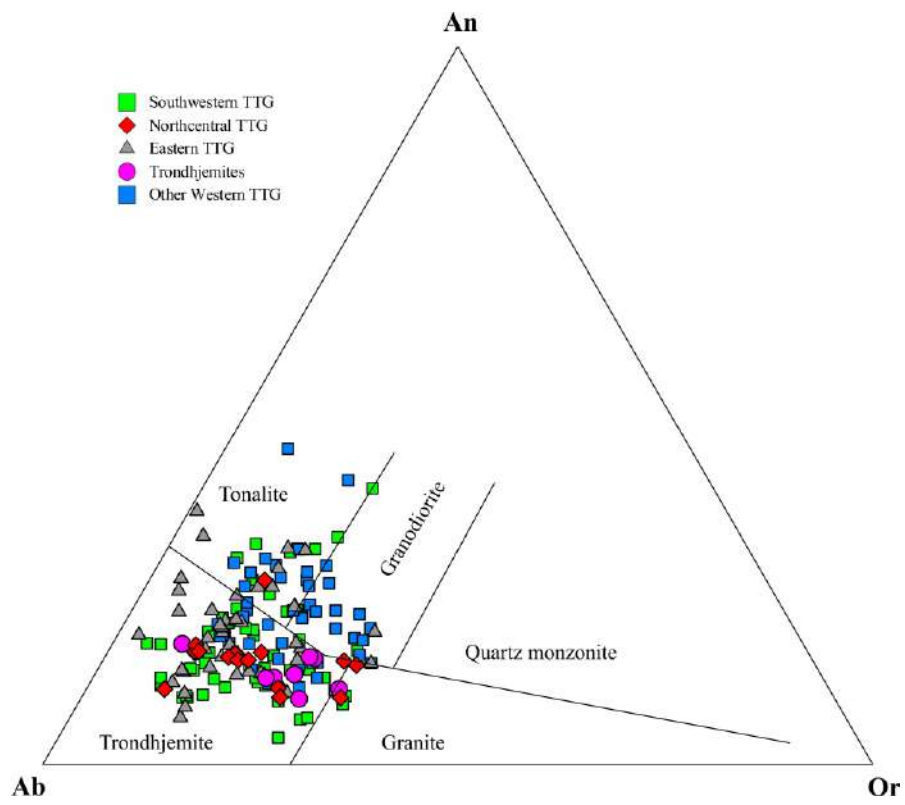


Fig. 7.9a

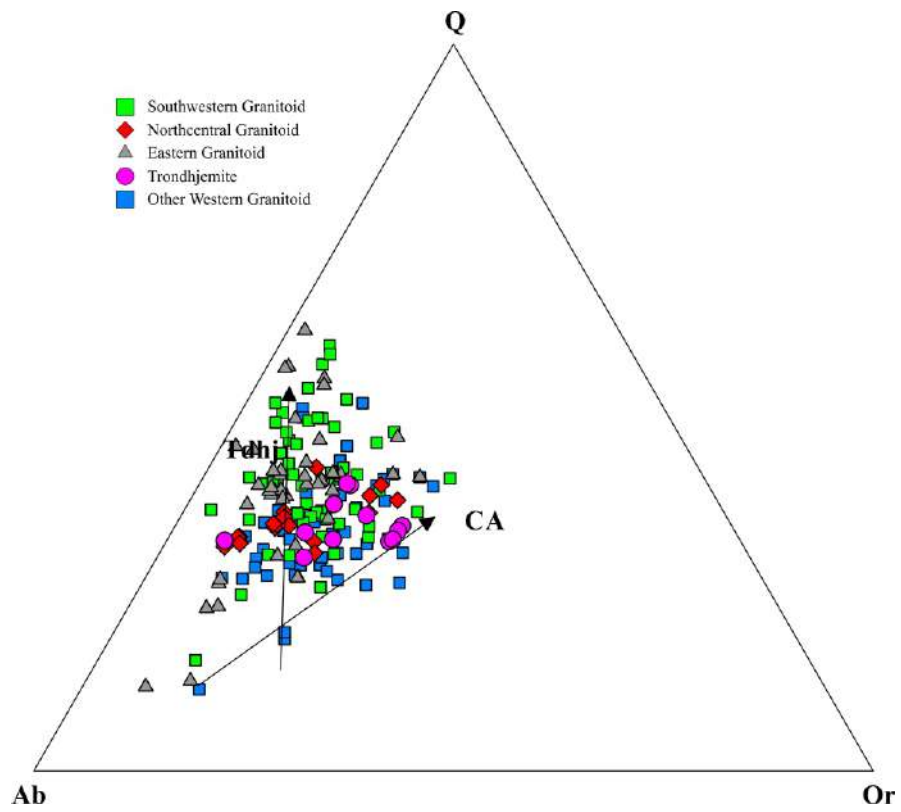


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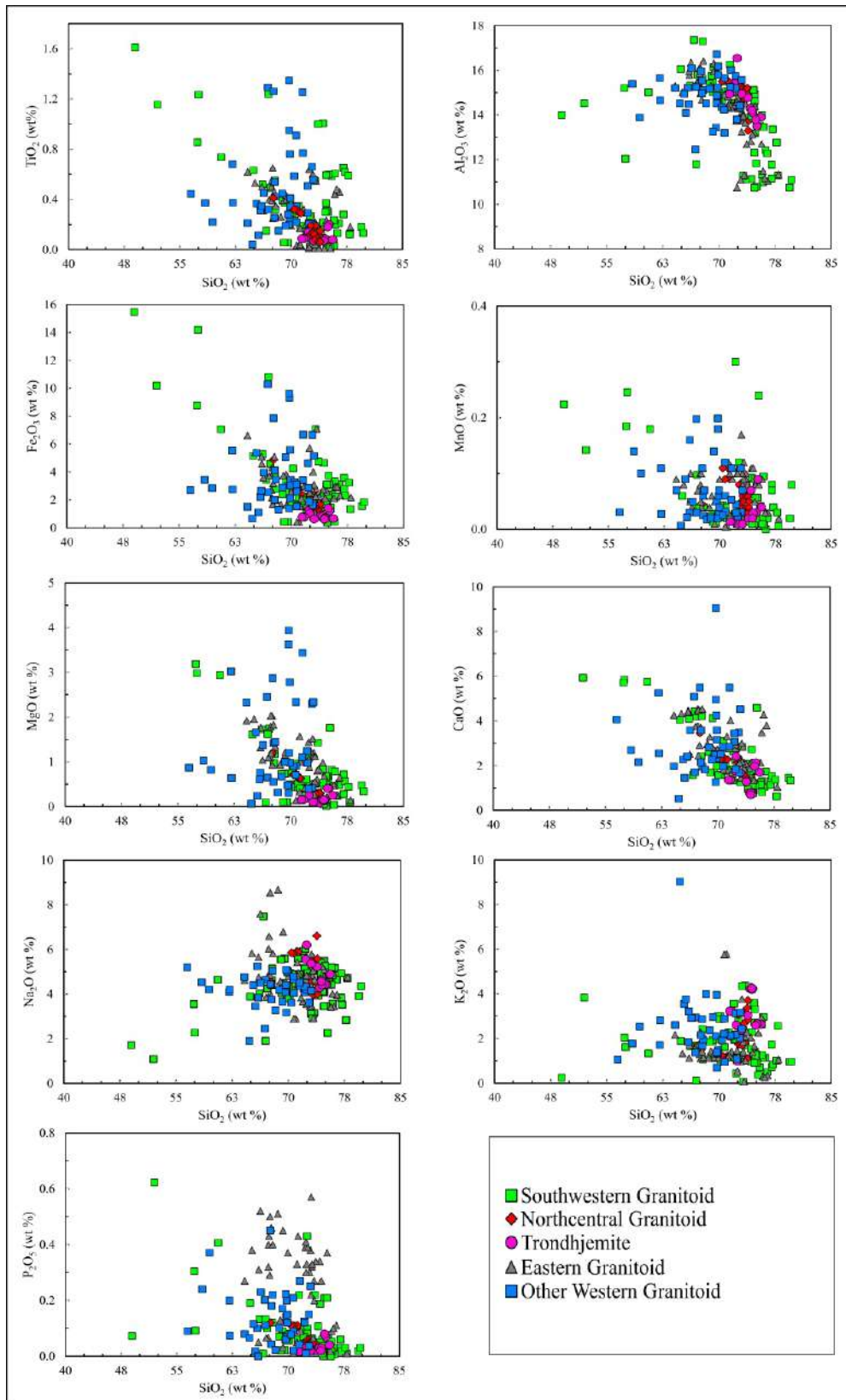


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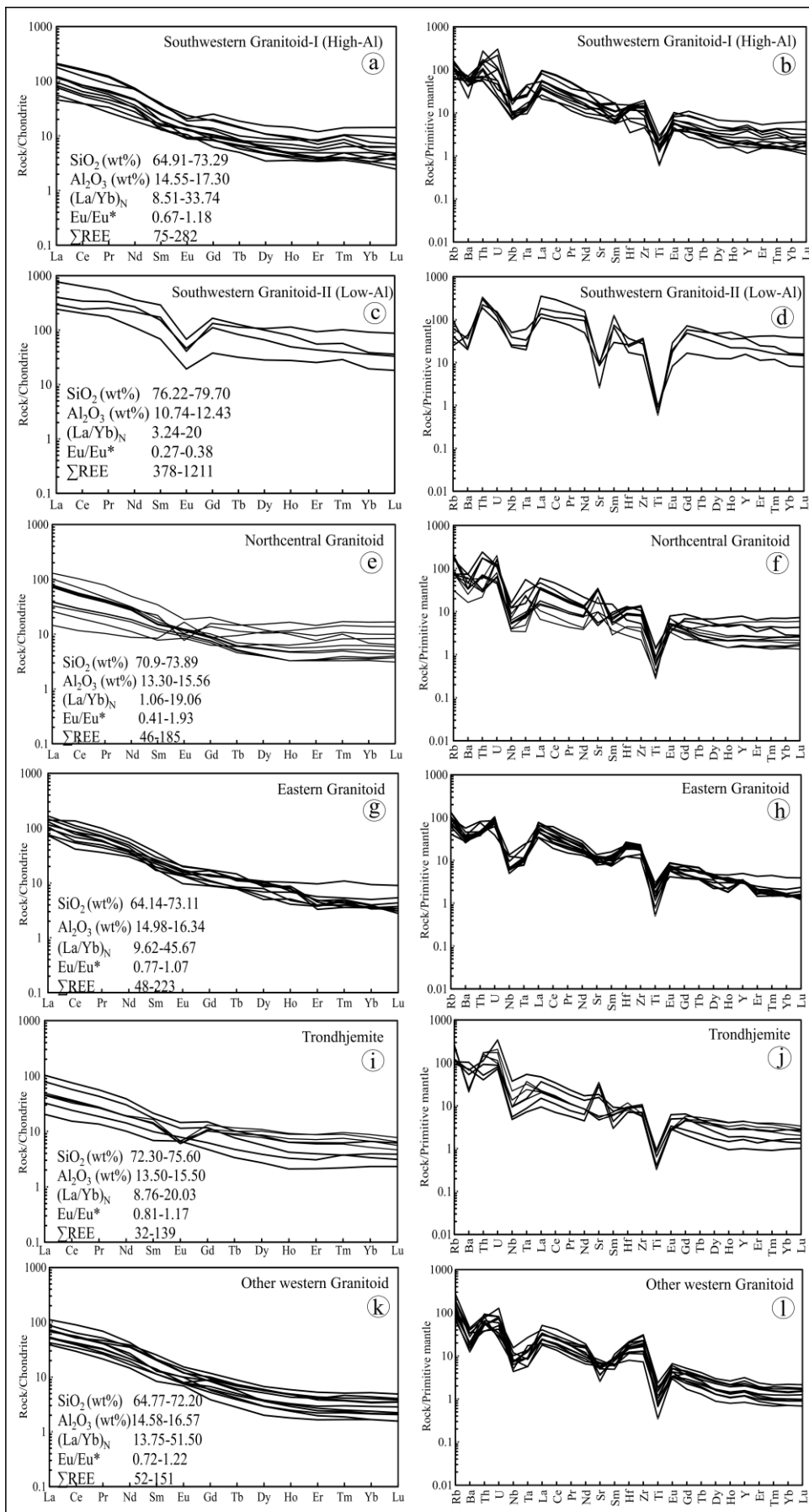


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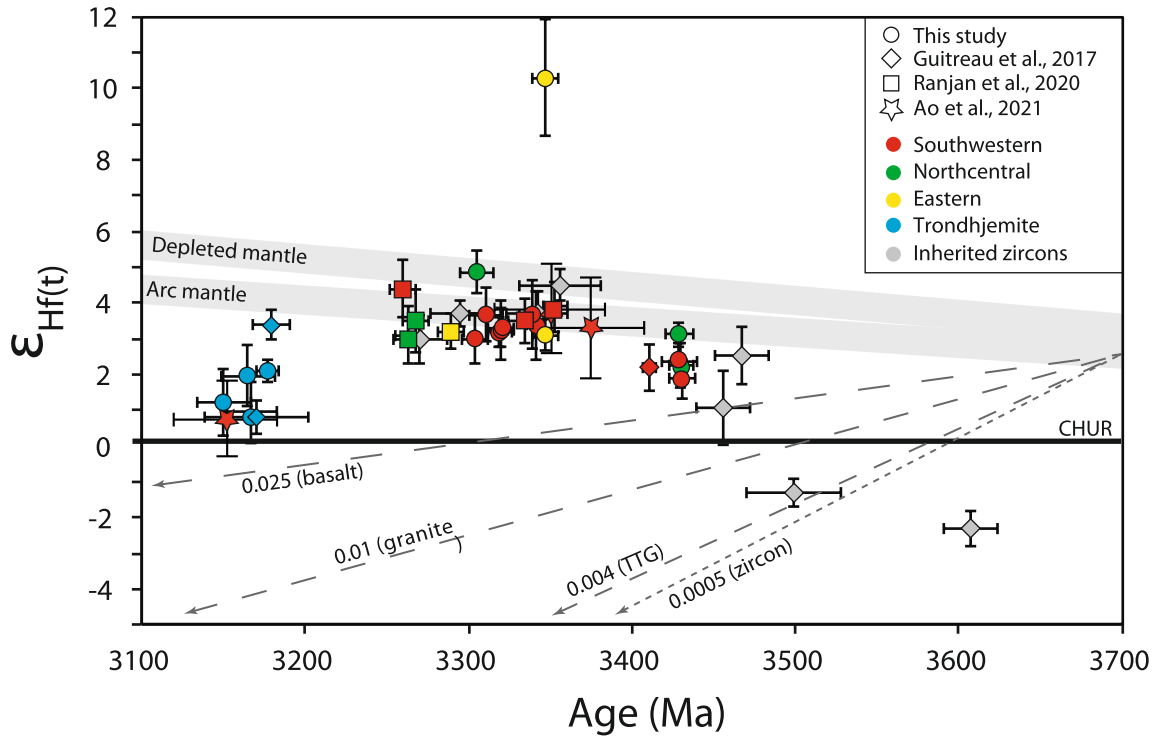


Fig. 7.11a

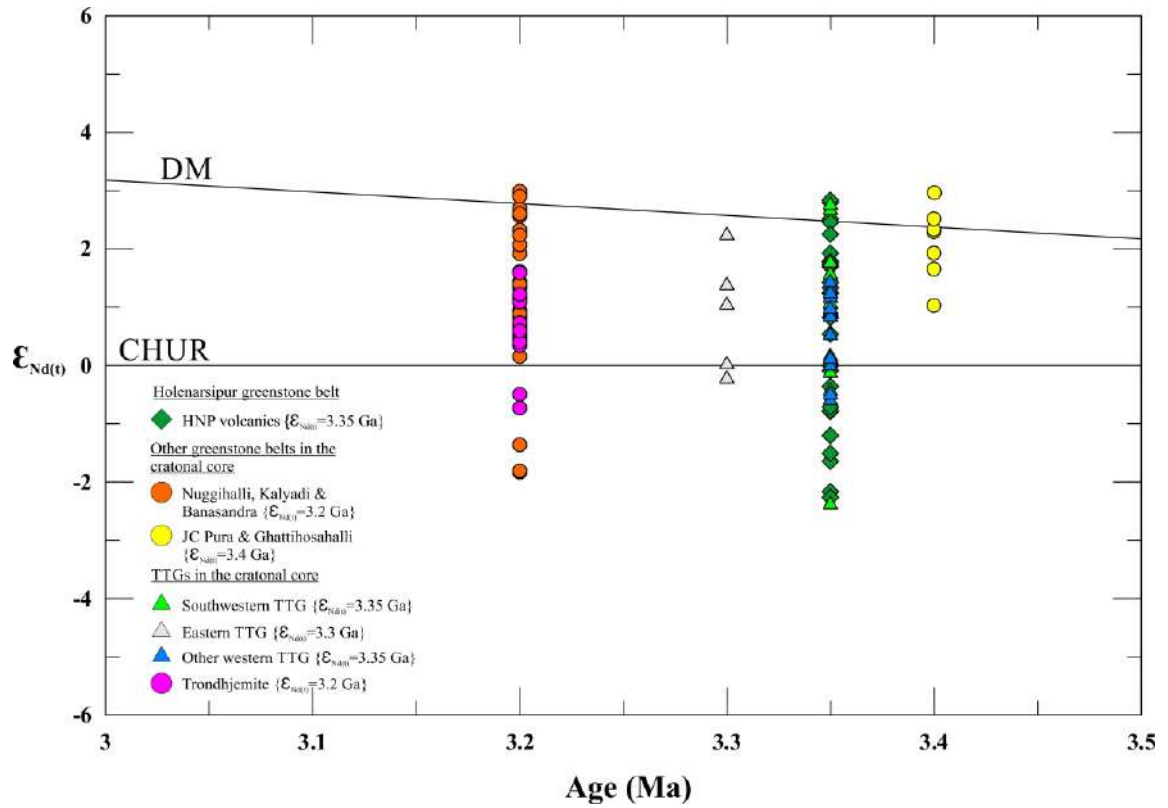


Fig. 7.11b

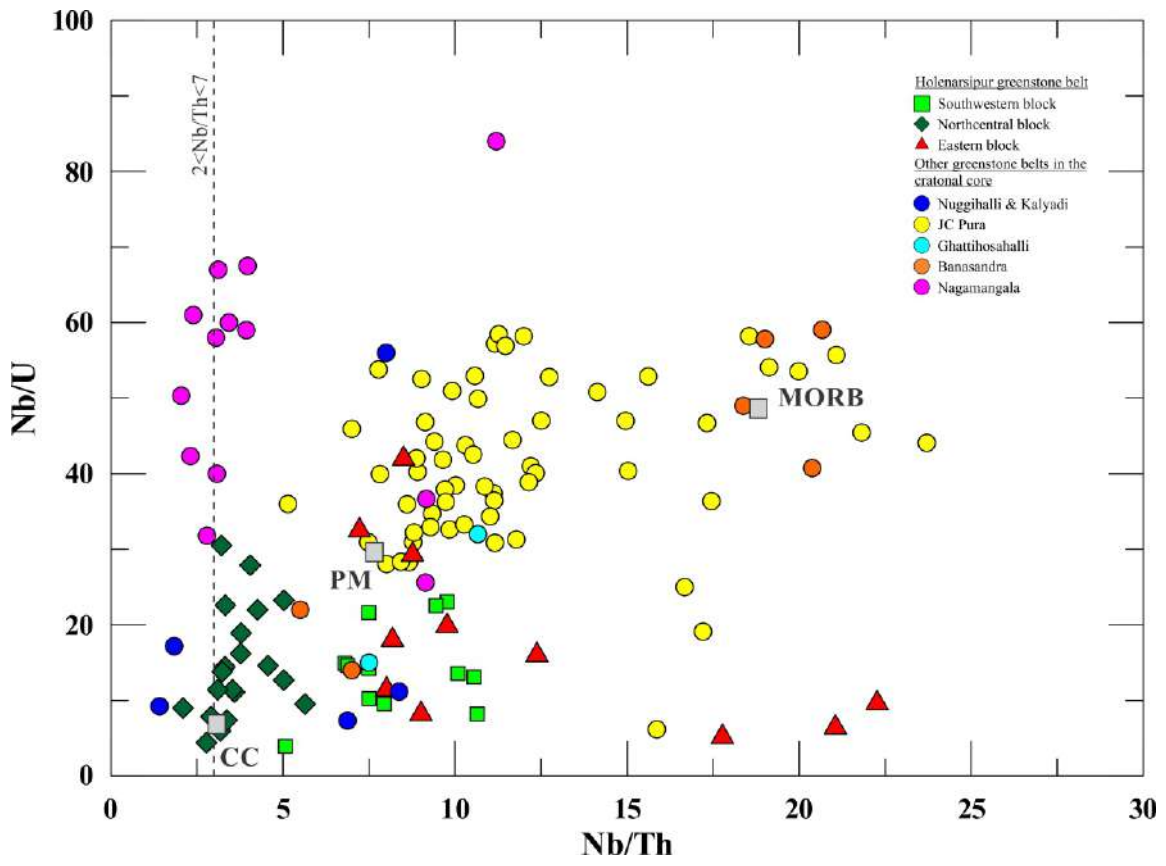


Fig. 8a

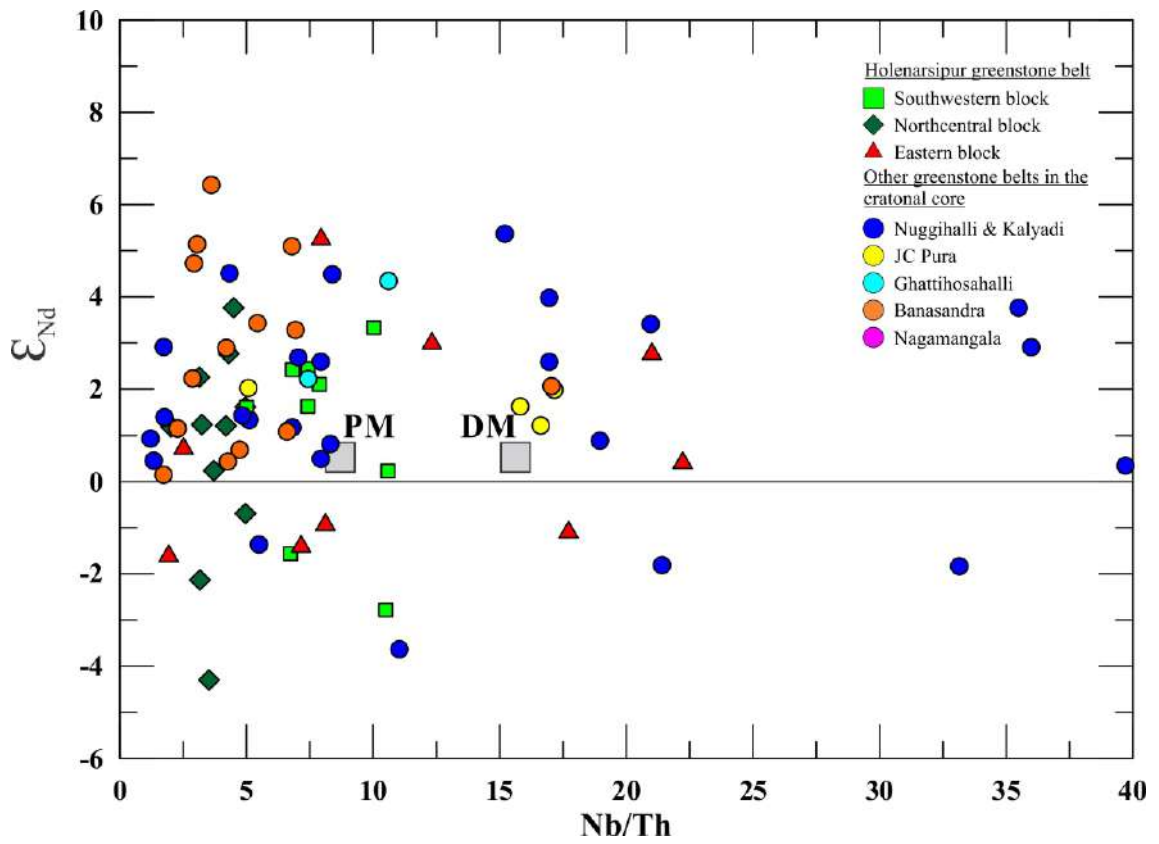


Fig. 8b

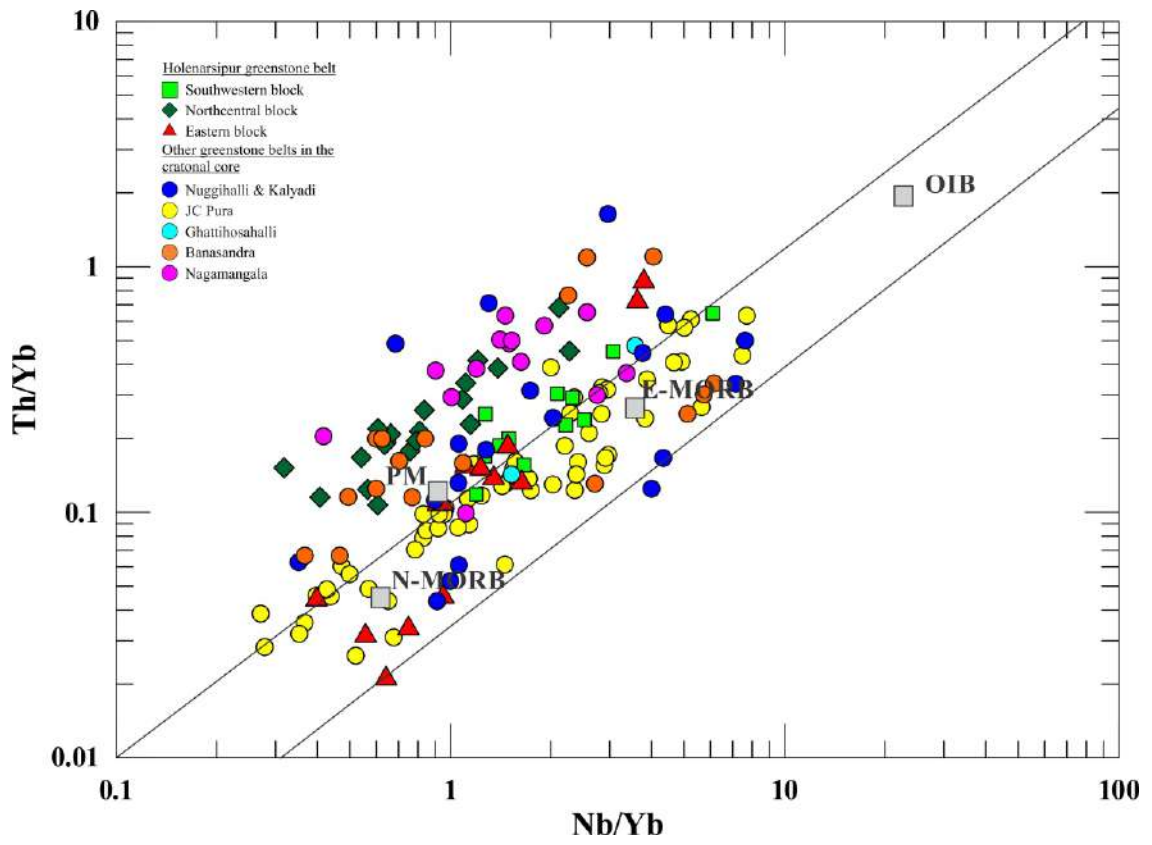


Fig. 8c

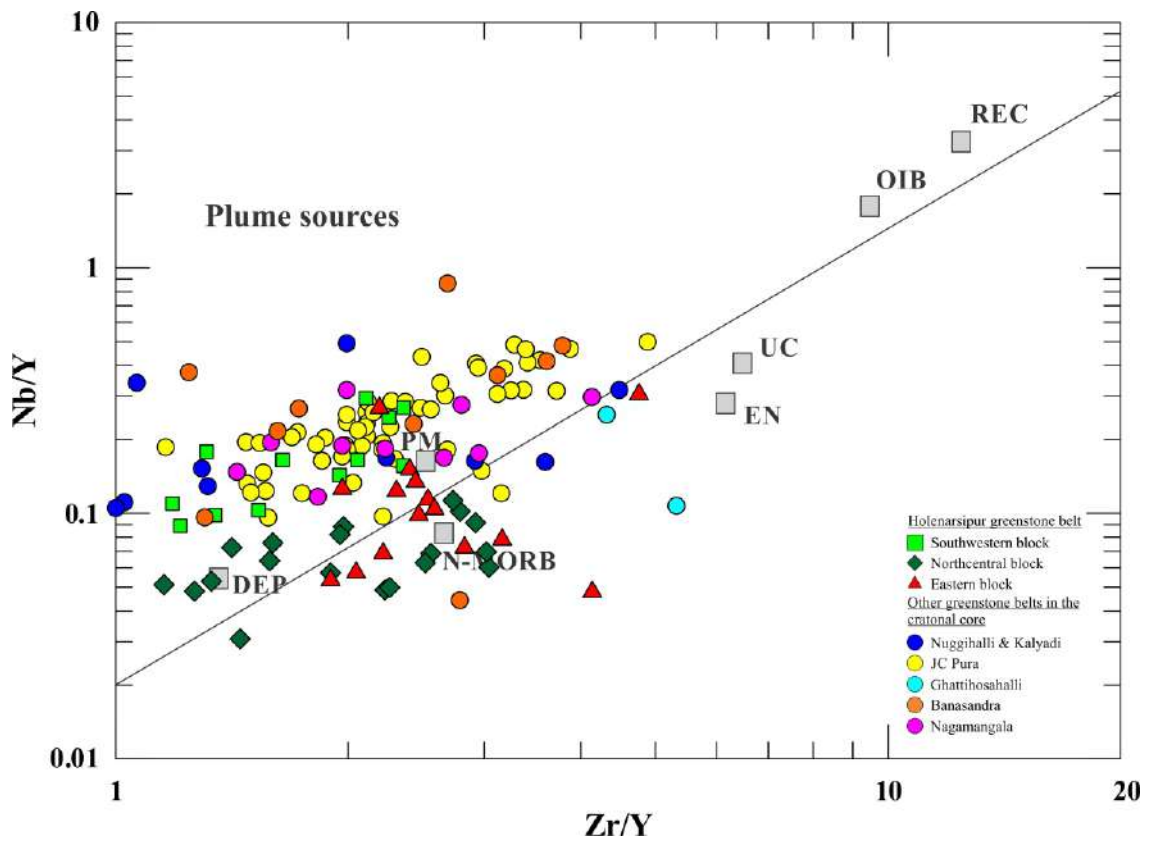


Fig. 8d

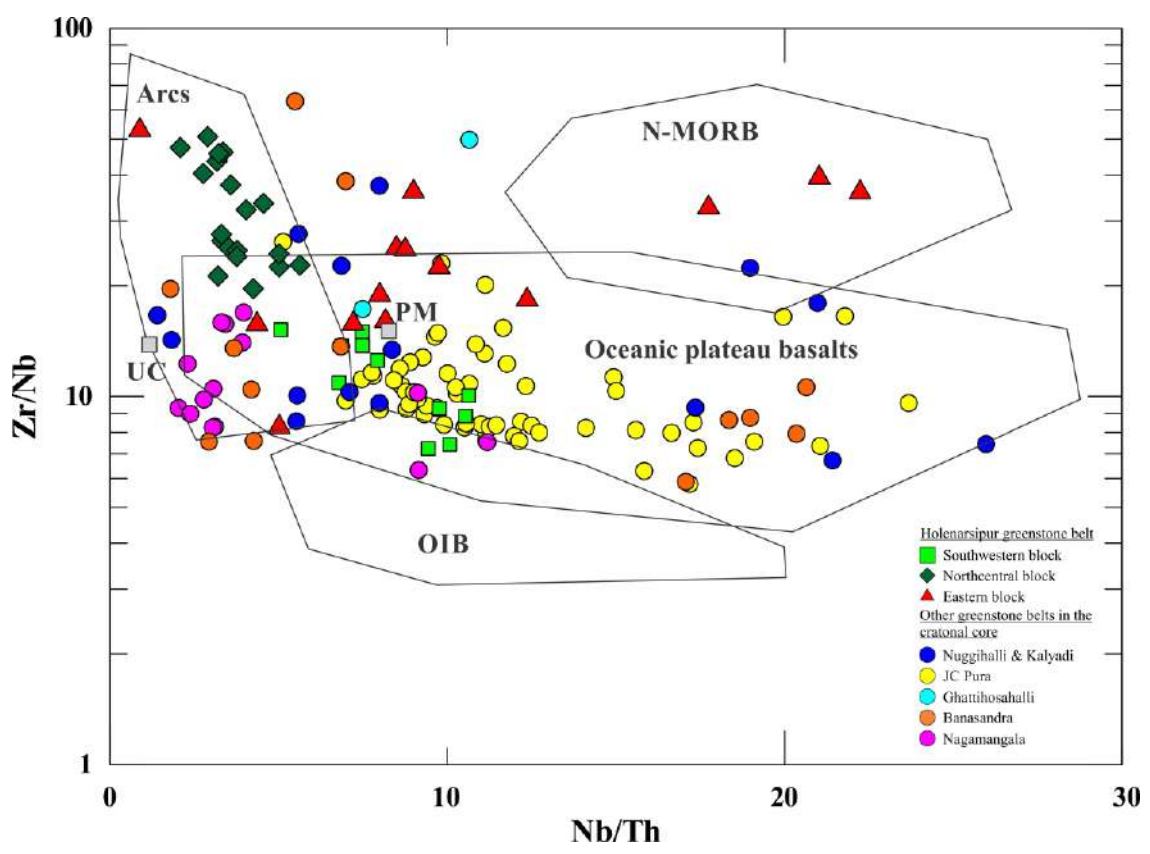


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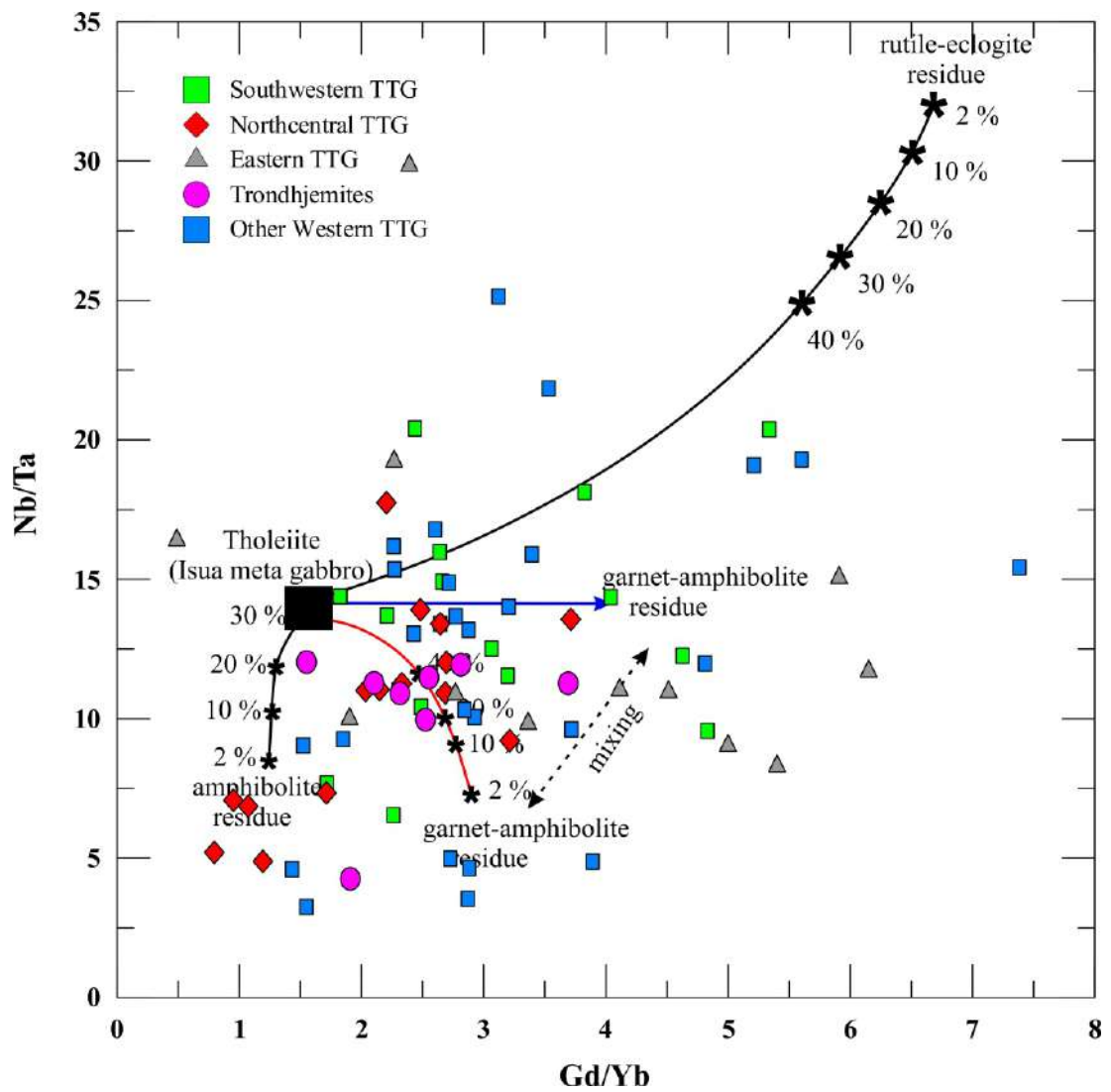


Fig. 9a

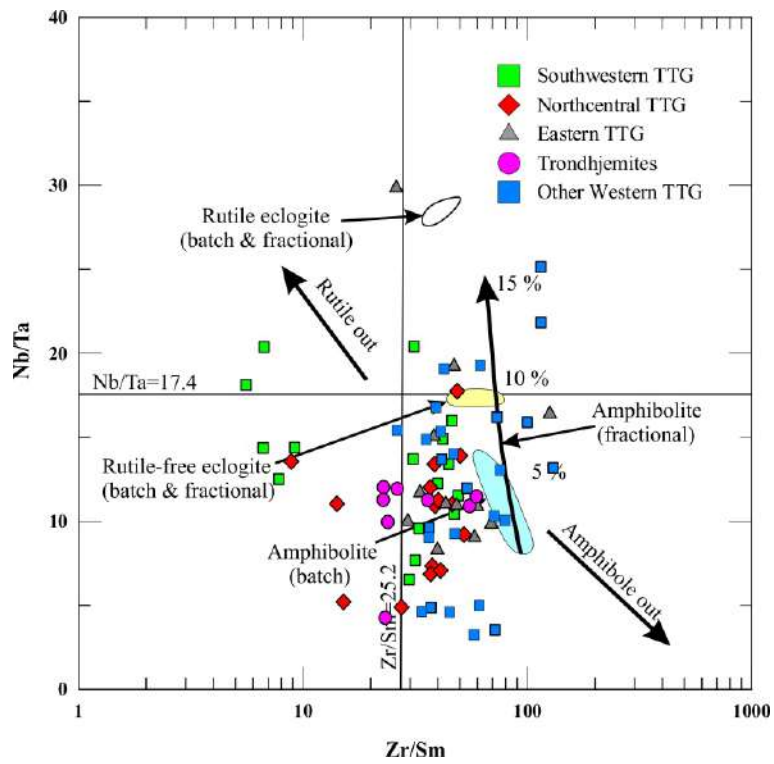


Fig. 9b

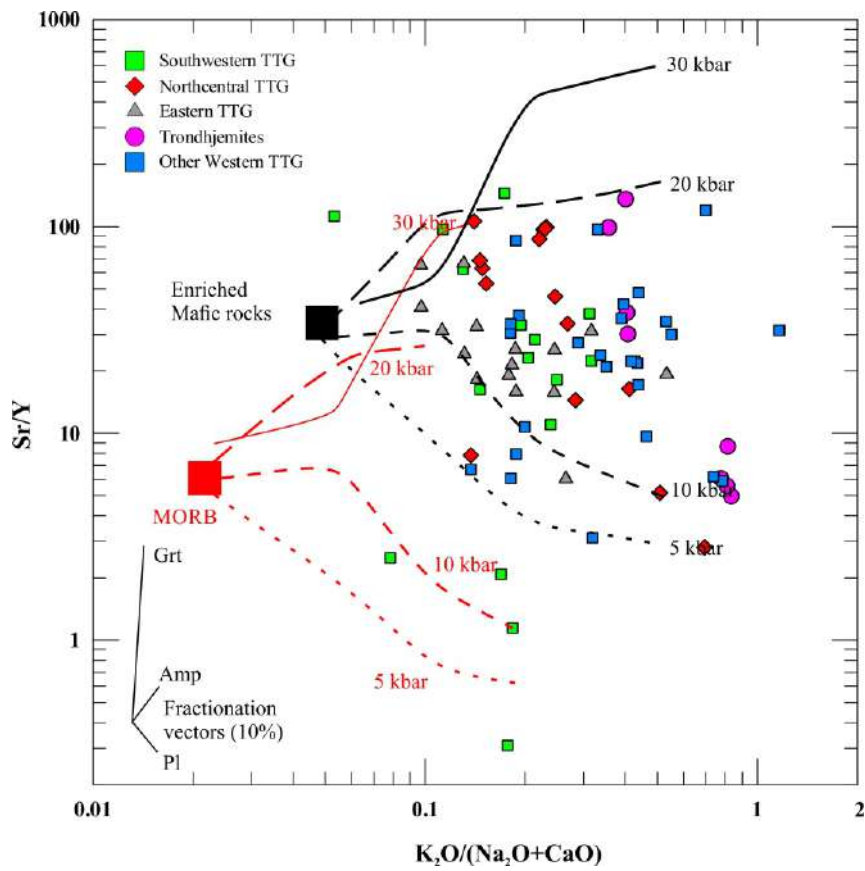


Fig. 9c

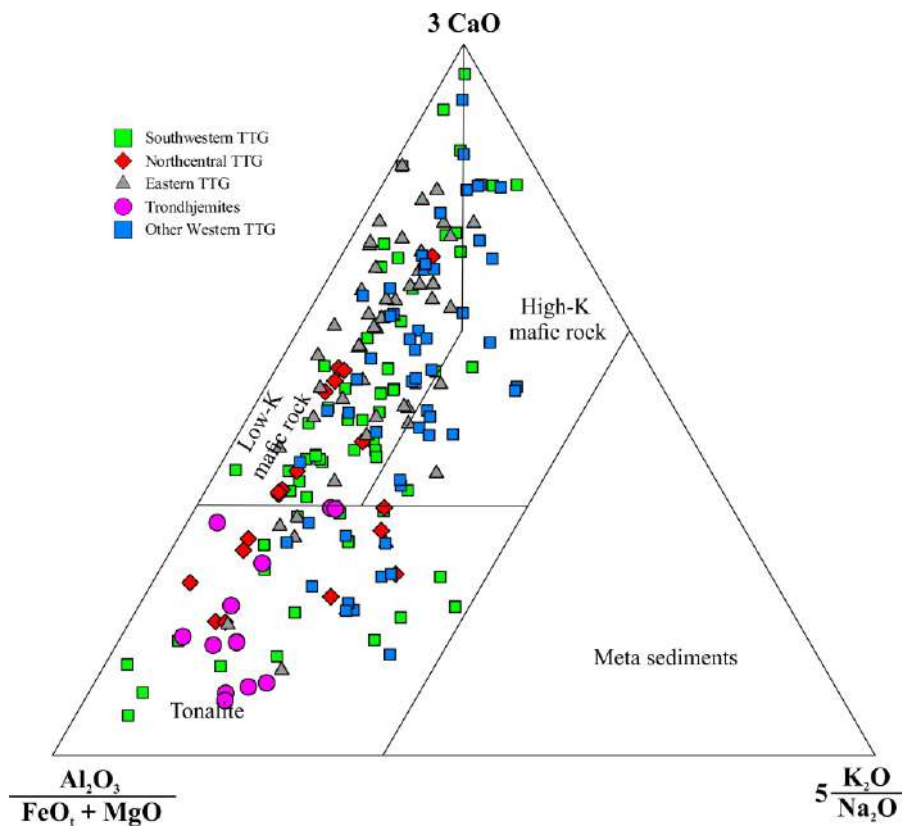


Fig. 9d

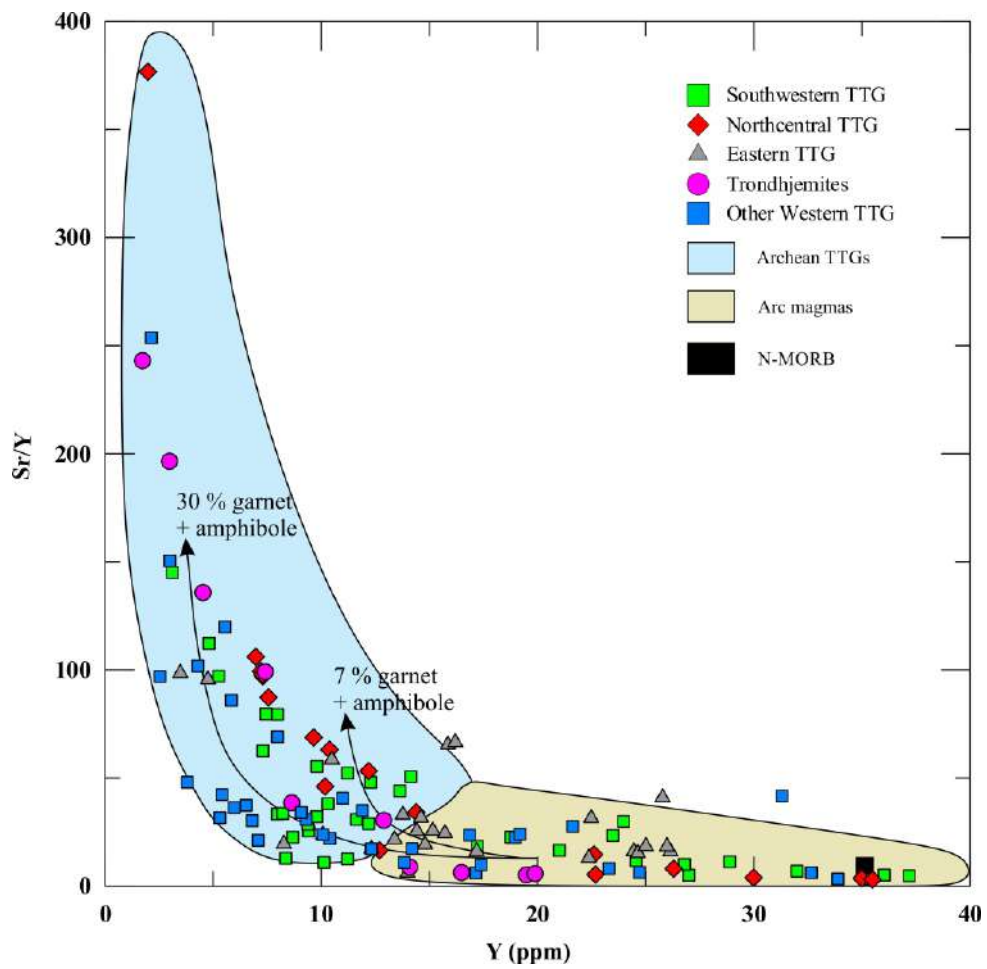


Fig. 9e

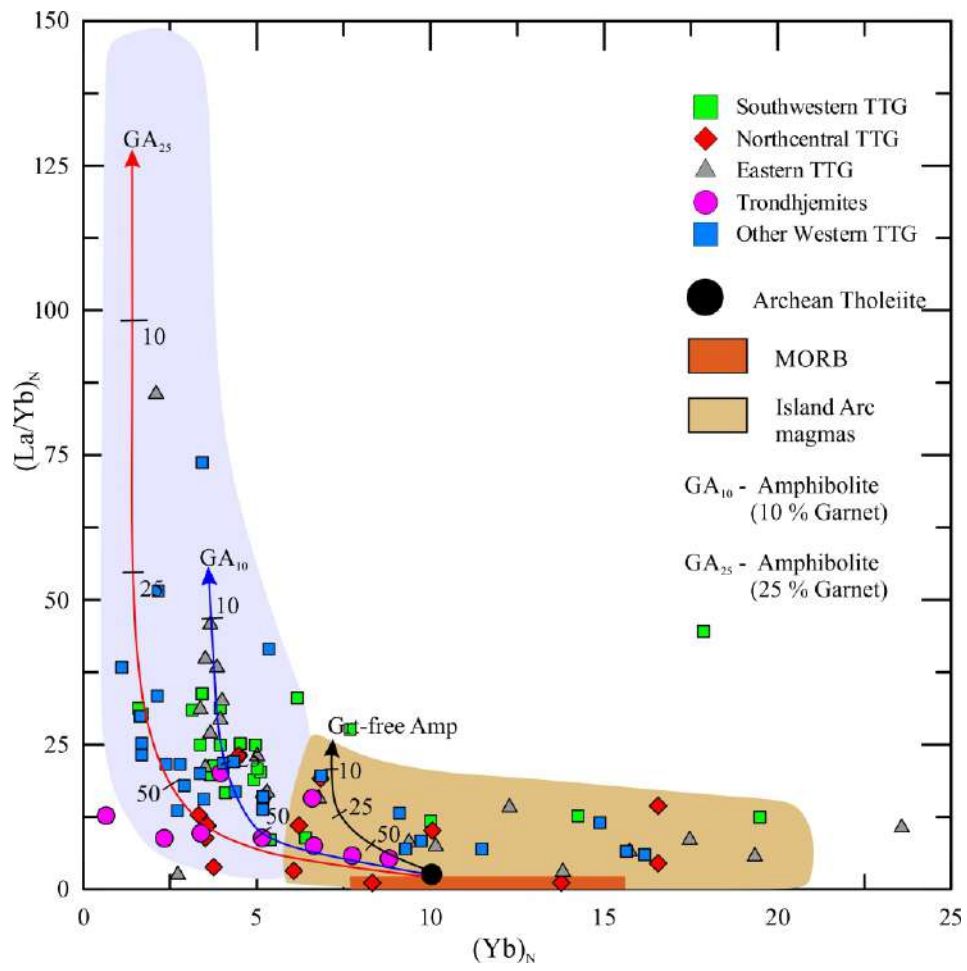


Fig. 9f