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Multi-stage crustal growth and Neoproterozoic geodynamics in the Eastern Dharwar Craton, southern India

M. Jayananda, K.R. R Aadhiseshan, Monika A Kusiak, Simon A Wilde,
Kowete-U Sekhamo, Martin Guitreau, M. Santosh, R.V. V Gireesh

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1 ***GR Focus Review***

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5 3 **Multi-stage crustal growth and Neoproterozoic geodynamics in the Eastern**
6 4 **Dharwar Craton, Southern India**
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24 Abstract

1
2 25 The Dharwar Craton is a composite Archean cratonic collage that preserves important records
3 26 of crustal evolution on the early Earth. Here we present results from a multidisciplinary study
4 27 involving field investigations, petrology, zircon SHRIMP U-Pb geochronology with *in-situ*
5 28 Hf isotope analyses, and whole-rock geochemistry, including Nd isotope data on migmatitic
6 29 TTG (tonalite-trondjemite-granodiorite) gneisses, dark grey banded gneisses, calc-alkaline
7 30 and anatectic granitoids, together with synplutonic mafic dykes along a wide Northwest –
8 31 Southeast corridor forming a wide time window in the Central and Eastern blocks of the
9 32 Dharwar Craton. The dark grey banded gneisses are transitional between TTGs and calc-
10 33 alkaline granitoids, and are referred to as ‘transitional TTGs’, whereas the calc-alkaline
11 34 granitoids show sanukitoid affinity. Our zircon U-Pb data, together with published results,
12 35 reveal four major periods of crustal growth (ca. 3360-3200 Ma, 3000-2960 Ma, 2700-2600
13 36 Ma and 2570-2520 Ma) in this region. The first two periods correspond to TTG generation
14 37 and accretion that is confined to the western part of the corridor, whereas widespread 2670-
15 38 2600 transitional TTG, together with a major outburst of 2570-2520 Ma juvenile calc-
16 39 alkaline magmatism of sanukitoid affinity contributed to peak continental growth. The
17 40 transitional TTGs were preceded by greenstone volcanism between 2746-2700 Ma, whereas
18 41 the calc-alkaline magmatism was contemporaneous with 2570-2545 Ma felsic volcanism.
19 42 The terminal stage of all four major accretion events was marked by thermal events reflected
20 43 by amphibolite to granulite facies metamorphism at ca. 3200 Ma, 2960 Ma, 2620 Ma and
21 44 2520 Ma. Elemental ratios [(La/Yb)_N, Sr/Y, Nb/Ta, Hf/Sm] and Hf-Nd isotope data suggest
22 45 that the magmatic protoliths of the TTGs emplaced at different time periods formed by
23 46 melting of thickened oceanic arc crust at different depths with plagioclase + amphibole
24 47 ±garnet + titanite/ilmenite in the source residue, whereas the elemental (Ba-Sr, [(La/Yb)_N,
25 48 Sr/Y, Nb/Ta, Hf/Sm]) and Hf-Nd isotope data [$\epsilon\text{Hf}_{(T)} = -0.67$ to 5.61 ; $\epsilon\text{Nd}_{(T)} = 0.52$ to 4.23 ;
26 49 of the transitional TTGs suggest that their protoliths formed by melting of composite sources
27 50 involving mantle and overlying arc crust with amphibole + garnet + clinopyroxene ±
28 51 plagioclase + ilmenite in the residue. The highly incompatible and compatible element
29 52 contents (REE, K-Ba-Sr, Mg, Ni, Cr), together with Hf and Nd isotope data [$\epsilon\text{Hf}_{(T)} = 4.5$ to -
30 53 3.2 ; $\epsilon\text{Nd}_{(T)} = 1.93$ to -1.26]; of the sanukitoids and synplutonic dykes suggest their derivation
31 54 from enriched mantle reservoirs with minor crustal contamination. Field, elemental and
32 55 isotope data [$\epsilon\text{Hf}_{(T)} = -4.3$ to -15.0 ; $\epsilon\text{Nd}_{(T)} = -0.5$ to -7.0] of the anatectic granites suggest
33 56 their derivation through reworking of ancient as well as newly formed juvenile crust. Secular
34 57 increase in incompatible as well as compatible element contents in the transitional TTGs to
35 58 sanukitoids imply progressive enrichment of Neoproterozoic mantle reservoirs, possibly through
36 59 melting of continent-derived detritus in a subduction zone setting, resulting in the
37 60 establishment of a sizable continental mass by 2700 Ma, which in turn is linked to the
38 61 evolving Earth. The Neoproterozoic geodynamic evolution is attributed to westward convergence
39 62 of hot oceanic lithosphere, with continued convergence resulted in the assembly of micro-
40 63 blocks, with eventual slab break-off leading to asthenosphere upwelling caused extensive
41 64 mantle melting and hot juvenile magma additions to the crust. This led to lateral flow of hot
42 65 ductile crust and 3D mass distribution and formation of an orogenic plateaux with subdued
43 66 topography, as indicated by strain fabric data and strong seismic reflectivity along an E-W
44 67 crustal profile in the Central and Eastern blocks of the Dharwar Craton.

45 68
46 69 Key words: Dharwar Craton; Zircon U-Pb geochronology; Lu-Hf-Sm-Nd isotopes; Crustal
47 70 evolution; Archean Earth.

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| 49 | 109 1. Introduction |
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| 51 | 110 The Archean witnessed major periods of juvenile crustal accretion, continental growth and |
| 52 | 111 craton formation (Goodwin, 1996; Hawkesworth et al., 2010; Dhuime et al., 2012; Condie et |
| 53 | 112 al., 2014). Models proposed for continental growth suggest that 60-70 percent of the |
| 54 | 113 preserved continental crust on our globe formed by 2500 Ma (e.g. Armstrong and Harmon, |
| 55 | 114 1981; Reymer and Schubert, 1984; Cawood et al., 2013; Hawkesworth et al., 2013; Taylor |
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115 and McLennan, 1995; Dhuime et al., 2018). During the past two decades, attention has been
116 focused on the changes in petrogenetic, thermal and tectonic processes involved in the
117 differentiation of crust from the mantle and the transformation of newly-formed juvenile crust
118 into Archean cratons (e.g. Martin et al., 2005; Mueller et al., 2010; Chardon et al., 2009;
119 Moyen, 2011; Spencer et al., 2017; Jayananda et al., 2018; Korenaga, 2018). The Archean
120 cratons were constructed by episodic juvenile accretion, recycling of pre-existing crustal
121 materials and assembly of micro-blocks through multiple collisions (e.g. Manikyamba et al.,
122 2017; Roberts and Santosh, 2018; Jayananda et al., 2018; Li et al., 2018a). However, the
123 tectonic context of crustal accretion remains equivocal, with debates on stagnant lid, through
124 vertical accretion associated with plumes, to modern-style horizontal tectonic processes
125 (Choukroune et al., 1997; Smithies et al., 2005; Bedard, 2006, 2013; Dhuime et al 2012;
126 Gerya, 2014; Sizova et al., 2015; Kamber et al., 2015; Brown and Johnson, 2018; Cawood et
127 al., 2018; Jayananda et al., 2018; Moyen and Laurent, 2018). Despite significant advances in
128 our understanding of the timing and mechanisms of major Archean craton building processes,
129 secular changes in the compositions of juvenile additions, as well as the tectonic context of
130 Archean continental growth, are still topics of debate. In this context, the Dharwar Craton
131 forms an ideal target as it consists of a large tilted panel of pristine Archean crust preserving
132 a wide time window (3600-2500 Ma) to address coupled evolution of the crust-mantle
133 system, secular changes in petrogenetic and geodynamic processes and craton formation.
134 Several integrated studies have contributed significantly to our understanding of the
135 fundamental architecture, time-frame of crustal accretion and reworking, as well as
136 petrogenetic processes (Chardon et al., 2008, 2011; Dey et al., 2016; Jayananda et al.,
137 2013a,b, 2015; Guitreau et al., 2017; Manikyamba et al., 2017; Wang and Santosh, 2019 and
138 references therein). More recent studies have shown that the Dharwar Craton formed by
139 assembly of micro-blocks with independent crustal histories (Jayananda et al., 2018; Roberts
140 and Santosh, 2018; Li et al., 2018a). However, the role of juvenile accretion and ancient
141 crustal reworking in Archean continental growth and craton building processes in individual
142 blocks is not precisely known, as most of the studies focussed on either greenstone volcanism
143 (Manikyamba and Kerrich, 2012; Jayananda et al., 2013b; Khanna et al., 2014, 2016; Yang et
144 al., 2015; Dey et al., 2018) or surrounding granitoids (Dey et al., 2012, 2014, 2015).
145 However, integrated studies on the age zonation patterns of the gneisses and granitoids, as
146 well as their spatial/temporal relationship with volcanism in the adjoining greenstone belts
147 and crust-mantle evolution through the Neoproterozoic, are rather limited. Furthermore, the
148 geodynamic context of crust formation and craton building processes are hotly debated,

149 especially between horizontal tectonics involving westward convergence of hot orogenic
150 crust (Chardon et al., 2011; Dey et al., 2016) or eastward subduction (Bidyananda et al.,
151 2016; Santosh and Li, 2018), or even plume (or combined plume-arc) models (Jayananda et
152 al., 2013a, 2018; Manikyamba et al., 2012). Consequently, the main purpose of this
153 contribution is to address the timing and relative roles of juvenile crust addition versus
154 reworking of Archean continental crust, the evolving petrogenetic and geodynamic processes
155 and craton building through integrated field and petrographic studies, U-Pb zircon
156 geochronology together with in-situ Hf isotope analyses, whole-rock geochemistry and Nd
157 isotopes. For this, we selected basement gneisses and granitoids along a wide corridor in the
158 Central and Eastern blocks of the Dharwar Craton from Hangal in the northwest to
159 Madanapalle to the southeast (Fig. 1), corresponding to different crustal levels.

2. Regional geological framework of the Dharwar Craton

161 The Dharwar Craton (Fig. 1) has been traditionally divided into two cratonic blocks, namely
162 the western and eastern craton, based on lithological associations, metamorphic grade and
163 crustal thickness (Swami Nath et al., 1976; Gupta et al., 2003; Jayananda et al., 2006;
164 Chardon et al., 2008). The boundary between these two crustal blocks was marked by the
165 steep mylonitic zone along the eastern boundary of the Chitradurga greenstone belt.
166 However, recent studies consider that the craton is composed of three distinct blocks, namely
167 the Western, Central and Eastern based on the age of the basement rocks, distinct thermal
168 records and accretionary histories (Peucat et al., 2013; Jayananda et al., 2013a,2018;
169 Dey,2013; Li et al., 2018a,b). The Western block extends from the Arabian Sea to the eastern
170 boundary of the Chitradurga greenstone belt and contains the oldest nucleus with 3400-3200
171 Ma TTG-greenstone assemblages that record two major pre-2500 Ma thermal events at ca.
172 3100-3000 Ma and 2620 Ma and record local reworking that resulted in the emplacement of
173 high-K granites (Chardon et al., 2011; Jayananda et al., 2006, 2013a, 2015, 2019). The
174 Central Dharwar block extends from the Chitradurga boundary shear zone (see Fig. 1) to the
175 western boundary of the Kolar – Kadiri-Penakacherla – Kushtagi-Hungund greenstone belts,
176 and comprises a mix of ca. 3400-3000 Ma and 2700-2500 Ma crust that was affected by two
177 pre-2500 Ma thermal events at ca. 3200 Ma and 2620 Ma (Jayananda et al. 2011, 2013a).
178 The Eastern Dharwar block extends eastward from the above-mentioned greenstone belts to
179 the Cuddapah Basin and Eastern Ghats Mobile Belt and contains mainly 2700-2530 Ma crust
180 that was affected by only the 2500 Ma thermal event, which commonly overprinted the whole
181 of the Archean crust in the Dharwar Craton (Peucat et al., 2013; Jayananda et al., 2013a).

182 Overall, the Dharwar Craton (Fig. 1) is a composite collage of Archean crustal blocks that
183 form a large tilted panel of pristine crust that evolved during a wide time window (3800-2500
184 Ma; (Jayananda et al., 2018 and references therein). The major lithological assemblages
185 include different types of gneisses, two generations of greenstone sequences and calc-alkaline
186 to potassic granitoid plutons. Recent studies have shown that the Dharwar Craton formed by
187 assembly of several micro-blocks (Santosh et al., 2015; Jayananda et al., 2018; Roberts and
188 Santosh, 2018; Li et al., 2018a,b). The polyphase gneisses are the most voluminous
189 lithologies and show large variations in mineralogy, chemical composition, timing of
190 emplacement of their protoliths (3600-2600 Ma), indicating they were affected by multiple
191 thermal events (Balakrishnan et al., 1999; Chardon et al., 2011; Moyen et al., 2003;
192 Jayananda et al., 2013a, 2015, 2018; Dey et al., 2016; Peucat et al., 2013). The greenstone
193 sequences are divided into the older Sargur Group and younger Dharwar Supergroup (Swami
194 Nath et al., 1981). The Sargur Group greenstone sequences, together with interlayered
195 tonalite-trondhjemite-granodiorite (TTG) gneisses, form the basement for the younger
196 Dharwar Supergroup greenstones. The two stratigraphically distinct greenstone sequences are
197 separated by angular unconformities (Ramakrishnan et al., 1976; Venkata Dasu et al., 1991;
198 Chardon et al., 1996). The older Sargur Group is dominated by ultramafic-mafic volcanics of
199 komatiite lineage that are interlayered with quartzite-pelite-carbonate-BIF (Banded Iron
200 Formation) units. Whole-rock Sm-Nd isochrons for komatiites display ages in the range of
201 3384-3150 Ma (Jayananda et al., 2008; Maya et al., 2017), whereas the associated felsic
202 volcanics record SHRIMP U-Pb zircon ages of ca. 3300 Ma (Peucat et al., 1995). Detrital
203 zircons from quartzites and pelites indicate ages in the range of ca. 3600-3230 Ma (Nutman et
204 al., 1992; Hokada et al., 2013; Lancaster et al., 2015; Wang and Santosh, 2019). In contrast,
205 the younger Dharwar Supergroup comprises ca. 3000-2740 Ma old greenstone sequences
206 unconformably overlying the basement (TTG/Sargur Group) and forming the vast
207 Bababudan-Chitradurga-Dharwar basins in the western part of the craton (Fig. 1). These
208 greenstone sequences are dominated by mafic to felsic volcanics with interlayered
209 conglomerate-quartzite-carbonate-greywacke-argillite-BIF sequences (Chadwick et al., 1981,
210 1985; Srinivasan and Ojakangas, 1986; Kumar et al., 1996; Jayananda et al., 2013b). The
211 greenstone basins in the eastern part of the craton with dominant volcanic sequences (Kolar-
212 Kadiri; Ramagiri-Penakacherla; Sandur; Gadwal-Narayanapet; Jonnagiri; Veligallu) show
213 bimodal age distributions at ca. 2750-2670 Ma and 2580-2545 Ma (Naqvi et al., 2002;
214 Manikyamba and Kerrich, 2012; Manikyamba et al., 2004, 2008, 2009, 2014, 2017;
215 Jayananda et al., 2013b; Dey et al., 2015,2018; Khanna et al., 2014,2016). Calc-alkaline to

216 potassic granitoids mark the terminal events of craton building in the Dharwar Craton. At
217 least three major episodes (ca. 3000 Ma, 2620 -2580 Ma and 2560-2520 Ma) of granitoid
218 emplacement are related to the craton forming events (Chardon et al., 2011; Jayananda et al.,
219 2006, 2018, 2019). The 3000 Ma high-K plutons intrude into TTG-Sargur Group greenstone
220 sequences in the Western Dharwar Craton (WDC; Chardon et al., 2011; Jayananda et al.,
221 2019) whereas the ca. 2620-2580 Ma potassic granites form discrete plutons which intrude
222 into either TTG basement or the basement-Dharwar Supergroup greenstone interface
223 (Jayananda et al., 2006; Chadwick et al., 2007; Sarma et al., 2012; Mohan et al., 2014). In
224 contrast, the 2560-2520 Ma granitoid plutons are restricted to the Central and Eastern
225 Dharwar craton and are regional in extent, forming major juvenile crustal accretions and
226 marking the final stage of cratonization (Jayananda et al., 2013a, 2018; Anand et al., 2014;
227 Dey et al., 2014, 2016). These granitoids are composite plutons comprising mostly
228 voluminous sanukitoids, with a minor component of anatectic granites (Peucat et al., 1993;
229 Harish Kumar et al., 2003; Jayananda et al., 1995, 2000; Balakrishnan et al., 1999; Moyen et
230 al., 2003a; Rogers et al., 2007; Dey et al., 2012, 2014, 2016).

231 On the cratonic scale, LANDSAT-TM data, together with our field-based fabric mapping,
232 reveal dome and basin patterns, steep north-trending foliations, flat foliations and regional
233 strike-slip shear zones (Chardon et al., 2008). The Western Dharwar Craton displays ca. 3200
234 Ma dome and basin patterns resulting from forces linked to inverse density stratification
235 associated with high-density greenstone loading on the low-density infracrustal basement
236 (Bouhallier et al., 1993, 1995; Chardon et al., 1996, 2008). The whole craton was involved in
237 Neoproterozoic crustal-scale regional shear deformation, reflected in widely-spaced shear zones
238 in the western part and closely-spaced shear zones in the eastern part. However, in the eastern
239 part (central and eastern blocks) conjugate shear zone patterns result from interference
240 between the two shear systems: a NE trending dextral shear system in the south and a sinistral
241 WNW trending shear system in the north (Chadwick et al., 2000, 2007; Chardon et al., 2008).
242 This shear deformation event is sub-contemporaneous with the emplacement of 2560-2520
243 Ma granitoid plutons and culminated with 2500 Ma regional metamorphism followed by slow
244 cooling until 2450 Ma (Chadwick et al., 2000; Chardon et al., 2002; 2008, 2011; Peucat et al.,
245 1993, 2013; Jayananda and Mahabaleswar, 1991; Jayananda et al., 2013a). The cratonized
246 Archean basement was later intruded by Paleoproterozoic to Mesoproterozoic mafic dyke
247 swarms (Srivastava et al., 2014a; Soderlund et al., 2018).

248 3. Geological setting of the study area

249 The studied crustal corridor (Hangal – Madanapalle) exposes a tilted NW-SE section of
250 preserved Archean crust that includes both the Central and Eastern blocks (Fig. 2). The major
251 lithological assemblages include different generations of gneisses with remnants of high-
252 grade supracrustal rocks, volcanic-sedimentary greenstone sequences and composite calc-
253 alkaline granitoid intrusions. In this study, the gneisses and granitoids of the Central and
254 Eastern blocks are described separately with their distinct thermal records and crustal
255 histories (Jayananda et al., 2013a).

3.1. Central block

257 The part of the studied corridor located in the Central block extends from the Chitradurga
258 boundary shear zone to the Kadiri-Kolar greenstone belt (Fig.2) and is composed of
259 migmatitic TTGs (Chardon et al., 2011; Jayananda et al., 2000, 2013a; Nasheeth et al., 2016;
260 Bidyananda et al., 2016), dark grey banded transitional TTGs, the Penukonda-Ramagiri-
261 Penakacherla greenstone belt and voluminous north-south trending composite granitoid
262 intrusions, including the Closepet granite, the Lepakshi-Bukkapatnam massifs and plutons
263 extending further east up to the Kadiri greenstone belt (Fig. 2). The migmatitic TTGs occur
264 as discontinuous outcrops and are abundant to the west of the Closepet granite. They are also
265 found as large screens along the boundary of the Closepet batholith (Fig. 3a) but are scanty
266 towards the eastern part of the central block. These migmatitic TTGs are weakly foliated and
267 comprise dark grey tonalite, grey granodiorite and whitish grey trondhjemite that commonly
268 is injected by granitoid veins or sheets. They are traversed by melt-filled dextral shear bands.
269 The TTGs are locally interlayered with high-grade supracrustal enclaves dominated by pelite-
270 quartzite-carbonate, with minor amphibolites. Migmatized pelitic rocks contain large
271 sillimanite crystals rimmed by spinel-quartz and traversed by melt-filled shears (Fig.3b).
272 Published detrital zircon data of the metapelites record ages as old as 3360 Ma for their
273 crustal provenance (Jayananda et al., 2013a). The P-T-t estimates, together with *in-situ*
274 monazite ages, reveal metamorphism close to ultra-high temperature (850°C at 5-7 kbar)
275 conditions at ca. 2620 Ma, with an earlier amphibolite facies event at ca. 3140 Ma (Jayananda
276 et al., 2011).

277 The dark grey banded gneisses are the most voluminous lithologies and they commonly
278 contain leucocratic veins with hornblende crystals developed along flat to shallow foliations.
279 The shallow foliations progressively become steep close to the large granitoid intrusions (e.g.
280 the Closepet batholith) and shear zones (Fig. 3c). Locally, melt-filled dextral shears traverse

281 dark grey banded gneisses that contain clinopyroxene crystals (Fig. 3d). Preliminary
282 petrographic and geochemical studies of the dark grey banded gneisses reveal their
283 transitional character between TTGs and sanukitoids and are therefore referred to as
284 ‘transitional TTGs’ (Jayananda et al., 2006, 2018; Dey et al., 2016). A dark grey hornblende-
285 rich quartz monzonite intrusion into migmatitic gneisses, known as the ‘Sira gneiss complex’
286 (Fig. 3e), was dated at 2562 ± 4 Ma by Chardon et al (2011).

287 The studied corridor in the Central block contains volcanic-sedimentary greenstone
288 sequences (e.g. Sandur, Penukonda- Ramagiri-Penakacherla greenstone belts) including
289 different types of metabasalts together with intermediate to felsic volcanics and minor
290 interlayered sediments. A Sm-Nd whole rock isochron for the metabasalts from the Sandur
291 greenstone belt records an imprecise age of 2705 ± 184 Ma (Naqvi et al., 2002), whereas
292 zircon U-Pb ages of the felsic volcanics in the higher stratigraphic levels record ages of 2691-
293 2658 Ma (Nutman et al., 1996). Zachariah et al (1995) reported Pb-Pb ages of 2746 ± 64 Ma
294 for metabasalts in the central part of the Ramagiri greenstone belt, whereas U-Pb zircon
295 dating of felsic volcanics from this belt records an age of 2707 ± 18 Ma (Balakrishnan et al.,
296 1999).

297 The granitoid intrusions in the Central block include the ~500 km long Closepet batholith,
298 which is the largest composite magmatic intrusion in the Dharwar Craton (Jayananda et al.,
299 1995, 2000, 2014; Moyen et al., 2003b) and the study area covers the central part of this
300 batholith (see Fig. 2) corresponding to the zone of accumulation (Moyen et al., 2003b). The
301 batholith contains two major suites of magmatic bodies; the more voluminous darker suite is
302 confined to the core, whereas the minor light pink to grey leucocratic anatectic granites are
303 located along the periphery, where they show diffusive contacts with the basement rocks.
304 The darker suite comprises dark grey quartz monzonite to monzodiorite (Fig. 3f) and
305 porphyritic monzogranite to granodiorite with large K-feldspar phenocrysts (30-40%) that
306 commonly show NNE to NNW orientation coinciding with the regional foliation (Fig. 3g).
307 The phenocrysts locally exhibit rapakivi-like texture, and rarely orbicular texture. The darker
308 suite contains fragmented synplutonic dykes and mafic magmatic enclaves (MME) that
309 commonly contain small feldspar phenocrysts. The MMEs and synplutonic mafic dykes
310 most likely correspond to late-stage mafic melts or hybrid melts injected into the crystallizing
311 host magma chamber wherein the MMEs form when mafic melts are introduced at the initial
312 stages of crystallization of the host (20-30 % crystals). The series of aligned MMEs and
313 fragmented synplutonic dykes formed when mafic melts entered into the host magma with

314 40-60 % crystals (Jayananda et al., 2009, 2014; Prabhakar et al., 2009; Gireesh et al., 2012).

315 Fragmented synplutonic dykes in the host granitoids have biotite-rich margins or show
316 magmatic flow fabrics along with their host. In some cases, close to the MMEs and
317 synplutonic mafic dykes, reversal of crystallization of the host granitoid resulted in back
318 veining that can be observed at the outcrop scale.

319 The light grey to pink anatectic granites occur as veins and large sheets and/or dykes. Locally,
320 gradation from gneisses to anatectic granite can be observed, but most commonly the granites
321 were injected into the gneisses (Fig. 3h) and intrusive relations suggest that these rocks formed
322 by anatexis of the basement at much deeper levels. The anatectic suite locally contains rotated
323 angular mafic to intermediate enclaves with strong fabrics (Fig. 3i) along the shear zones. The
324 anatectic rocks, together with synplutonic mafic dykes, define magmatic flow structures
325 implying injection of mafic melts with contemporaneous emplacement of anatectic magmas (Fig.
326 3j). Different stages of interaction between the host granitoid and mafic magma injections
327 suggest magma mixing and mingling.

328 *3.2. Eastern block*

329 The part of the studied corridor, located in the Eastern block, comprises dark grey banded
330 gneisses, voluminous migmatitic gneisses of diatexitic nature, volcanic-dominated greenstone
331 sequences (Kadiri and Veligallu) and granitoid intrusions (Fig. 2). The dark grey banded
332 gneisses occur as low outcrops with shallow foliations (Fig. 4a). Diatexitic gneisses are
333 widespread throughout the eastern block and they contain numerous biotite schlieren, mafic
334 xenoliths and screens of mafic gneisses (Fig. 4b). The diatexitic gneisses also show a
335 transition to banded grey gneisses and are locally interlayered with the dark grey banded
336 gneisses (Fig. 4c).

337 The greenstone sequences include the north-south trending Kadiri belt and the Veligallu
338 greenstone belt further east (see Fig. 2). The Kadiri greenstone belt contains mafic to felsic
339 volcanics with minor sediments. Zircon SHRIMP (Sensitive High Resolution Ion
340 Microprobe) U-Pb data for the felsic volcanics indicate a bimodal age distribution at $2717 \pm$
341 24 Ma and 2556 ± 13 , with inherited zircons recording ages of 2970 Ma and 3150 Ma also
342 present (Jayananda et al 2013b). Zircon from a rhyolite in the sequence recorded an age of
343 2535 ± 32 Ma (Dey et al., 2015). The Veligallu greenstone belt contains minor ultramafic
344 volcanic rocks with abundant mafic and felsic volcanics. Zircon SHRIMP U-Pb dating of the
345 felsic volcanics yielded an age of 2692 ± 5 Ma (Jayananda et al. 2013b) and a similar age of

346 2696 ± 54 Ma was obtained by a subsequent whole rock Lu-Hf isotope isochron on the
347 ultramafic volcanics (Khanna et al., 2016).

348 The granitoid plutons consist of quartz monzonite, monzodiorite, porphyritic monzogranite, and
349 granodiorite and are surrounded by anatectic pink to light grey granite and by voluminous
350 diatexites. Locally, angular fragments or boudins of quartz monzonite/monzodiorite and
351 porphyritic monzogranite are found in the diatexites (Fig. 4c). Dark grey hornblende-rich
352 monzodiorite occurs as discontinuous outcrops that contain mafic magmatic enclaves. The light
353 pink to whitish grey anatectic rocks commonly contains mafic enclaves, fragments of banded
354 grey gneisses and biotite schlieren. They locally show magmatic flow fabrics with fragments of
355 other granitoids (Fig. 4e). At several outcrops, the anatectic rocks contain MME. As an
356 example, a spectacular roadside outcrop (1 km south of Tanakallu along the Kadiri –
357 Madanapalle road) displays anatectic granite containing numerous MME /hybrid enclaves where
358 different stages of interaction, mixing, mingling and hybridization are visible (Fig. 4f). A large
359 quarry near Madanapalle exposes both intrusive granitoids and anatectic granites and diatexites
360 with numerous mafic enclaves where titanite crystal accumulation is observed in the anatectic
361 granite. Numerous northwest-trending quartz reefs traverse the granitoid massifs and these are
362 present only in the eastern block, probably related to the final differentiated silica-rich liquids
363 injected into the crystallized plutons.

364 The studied corridor displays spectacular flat to shallow fabrics which progressively become
365 steeper toward shear zones. Several N10-30°E trending melt-filled ductile shear zones, together
366 with N 10-40°W trending sinistral shear zones, define interference shear zone patterns
367 (Chadwick et al., 2000; Jayananda et al., 2006; Chardon et al., 2008, 2011). These interfering
368 shear zones are not confined to the margins of the granitoid plutons, but also cross-cut the
369 plutons that contain numerous angular mafic to ultramafic xenoliths, possibly transported from
370 the lower crust or mantle. In summary, the studied corridor exposes a tilted panel of Archean
371 continental crust with different lithological assemblages displaying shallow to steep fabrics and
372 with interfering shear zone patterns, thus forming a complex crustal architecture (Fig. 5)

373 4. Petrography

374 The migmatitic gneisses of the Central block are medium- to coarse-grained and contain quartz,
375 sodic plagioclase (An₁₂₋₁₅), minor microcline and biotite, with accessory zircon, apatite and
376 opaque Fe-Ti oxides (Fig. 6a). The banded dark grey to grey gneisses (transitional TTG) contain

377 quartz, more calcic plagioclase (An₁₅₋₂₈), microcline, abundant hornblende (Fig. 6b) and biotite,
378 with accessory zircon, apatite, titanite and opaque minerals.

379 The quartz monzonite contains quartz, plagioclase (An₁₅₋₂₆), abundant microcline, hornblende,
380 biotite (Fig. 6c), and local clinopyroxene, with accessory zircon, apatite, titanite, allanite and
381 opaque minerals. The porphyritic monzogranites are coarse-grained and display a bimodal grain
382 size with large alkali feldspar phenocrysts set in a quartz-feldspar matrix (Fig. 6d). The
383 accessory zircon, apatite, titanite, allanite and opaque minerals occur together and form clusters
384 with the mafic minerals. Plagioclase (An₁₅₋₁₈) shows crude zoning, whereas Potassium feldspar is
385 mainly microcline that is locally mantled by plagioclase, implying magma mingling.

386 The fragmented synplutonic dykes are fine- to medium-grained and contain small alkali feldspar
387 phenocrysts. The mineral assemblage consists of variable proportions of plagioclase (An₂₀₋₃₂)
388 and microcline and abundant hornblende and biotite, with some clinopyroxene and minor quartz
389 (Fig. 6e). Accessory minerals include zircon, large apatite needles, titanite and opaque minerals.

390 The anatectic granite is medium- to coarse-grained and has a hypidiomorphic granular texture.
391 These rocks contain abundant quartz and microcline, with subordinate plagioclase (An₁₀₋₁₂) and
392 biotite (Fig. 6f). Locally, the anatectic granites contain large crystals of hornblende that are
393 commonly replaced by biotite. The accessory minerals are zircon, apatite, titanite and opaques
394 that commonly occur together as clusters.

395 In the Eastern block, the banded gneisses are medium- to coarse-grained with a dominant
396 mineral assemblage of quartz, plagioclase (An₁₅₋₂₆), microcline, hornblende, and biotite, with
397 accessory zircon, apatite and opaque minerals (Fig. 7a). The plutons of quartz monzonite to
398 monzodiorite contain quartz, sub-equal plagioclase (An₁₅₋₂₈) and microcline, hornblende, biotite
399 (Fig. 7b), whereas the porphyritic varieties contain quartz, plagioclase (An₁₄₋₂₃), and abundant
400 microcline (Fig. 7c), with accessory zircon, apatite, titanite, and opaque minerals. The anatectic
401 granite and diatexite are medium- to coarse-grained with mafic minerals defining a magmatic
402 foliation. The mineral assemblage includes quartz, microcline, sodic plagioclase (An₁₂₋₁₅),
403 biotite, and local hornblende, with accessory zircon, titanite, and opaque minerals (Fig. 7d).

404 In summary, the intrusive rocks in the Central block are more voluminous and contain abundant
405 hornblende, with local clinopyroxene, when compared to the intrusive rocks in the Eastern block.
406 The anatectic rocks are more abundant in the Central block, whereas diatexites are more
407 common in the Eastern block.

408 5. Zircon U-Pb geochronology and *in situ* Hf isotopes

409 The SHRIMP U-Pb isotope data for zircon grains from migmatitic gneisses, transitional TTGs,
410 sanukitoids and anatectic granites analysed in this study are presented in [Supplementary table 1](#).
411 *In-situ* Hf isotope data of the corresponding zircons analysed for U-Pb are presented in
412 [Supplementary table 2](#). Details of the analytical methods are presented in the [Appendix](#).

413 5.1 Central block

414 5.1.1 Migmatitic and Transitional TTGs

415 *Sample EDC-TTG (14°39'10"N; 76°40'45"E)-Migmatitic tonalitic gneiss*

416 The tonalitic rocks in the migmatitic TTG were collected ~40 km north of Challakere town in
417 Bellary road (adjacent to telecom tower, see [Fig. 2](#)). Prismatic zircon crystals with pyramidal
418 terminations mostly reveal oscillatory zoning in cathodoluminescence (CL) images. Most grains
419 show discordant truncated growth zoning in cores, probably corresponding to dissolution.
420 Analyses were performed on prismatic grains with or without oscillatory zoning. They have low
421 U (129-481 ppm) and Th (62-462 ppm) contents, with Th/U ratios ranging from 0.10 to 1.10,
422 suggesting a magmatic origin. The results are discordant, with an upper intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age
423 of 3271 ± 5 Ma with MSWD = 1.3 ([Fig. 8.1a](#)), corresponding to the time of crystallization.

424 Zircons record $^{176}\text{Hf}/^{177}\text{Hf}$ ratios varying from 0.28073 to 0.28088, $\epsilon\text{Hf}_{(T)}$ values of +0.6 to +5.4
425 which were calculated using their corresponding SHRIMP zircon U-Pb age of 3271 Ma.
426 Depleted mantle model (T_{DM}) ages range between 3594-3321 Ma (T_{DM1}) and two-stage model
427 ages (T_{DM2}) between 3474-3301 Ma.

428 *Sample 17 CL-01 (13°52'32"N; 77°05'35"E)-Banded dark grey tonalitic gneiss*

429 This sample was collected along the western boundary of the Closepet batholith about 3 km
430 north of Rolla ([Fig. 2](#)). Zircon crystals from sample 17CL01 are elongated, euhedral, and most
431 exhibit well-defined oscillatory zoning typical of a magmatic origin. Some brighter domains
432 resembling metamorphic recrystallization can also be seen within a few of the analyzed crystals,
433 together with darker domains that likely result from local radiation-damage-induced
434 metamictization. No overgrowths or core-rim relationships are visible in the analyzed crystals.
435 The zircons contain low to high U (98-915 ppm) and Th (24-390 ppm) concentrations, with
436 Th/U ratios ranging from 0.11 to 0.57. A set of 24 analyses define an upper intercept $^{207}\text{Pb}/^{206}\text{Pb}$

age of 2670 ± 8 Ma (MSWD=1.50) corresponding to crystallization of the magmatic precursor of the tonalitic gneiss (Fig. 8.1b).

Sample KO-32: (13°47'52"N; 77°43'42"E)-Banded dark grey tonalitic gneiss

Sample KO-32 was collected ~3 km south of Lepakshi (Fig. 2). It is a coarse-grained, strongly banded dark grey tonalitic gneiss. CL images of zircon grains reveal elongated euhedral morphology with local recrystallized zones. They contain low to moderate concentrations of U (225-728 ppm) and Th (34-420 ppm, with a few outliers of very low Th <5 ppm) with Th/U ratios ranging from 0.16 to 1.21. A suite of euhedral zircons (12 analyses) is concordant to sub-concordant and defines an upper intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2601 ± 16 Ma (MSWD= 6.1, Fig. 8.1c).

Sample KO-20: (14°07'42"N; 77°41'11"E)-Banded dark grey tonalitic gneiss

Sample KO-20 was collected ~13 km west of Puttaparthi (Fig. 2) and is a banded tonalitic gneiss with a shallow foliation. Zircons are euhedral, elongated prismatic grains with aspect ratios of 3:1 or greater. They show concentric zoning typical of magmatic growth, with evidence of recrystallization of the edges locally observed in CL. Small inclusions of apatite are present. The zircons are characterized by moderate to high contents of U (300-1018 ppm) and Th (152-495 ppm, with few low values <16 ppm) and low Th/U ratios (0.24-0.55), indicative of a magmatic origin. A set of twenty-two zircons define an upper intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2600 ± 11 Ma. The weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age calculated from the concordant data is 2595 ± 15 (11/22 analyses, MSWD=8.9) (Fig. 8.1d).

In-situ Hf isotope analysis of zircons record $^{176}\text{Hf}/^{177}\text{Hf}_{(\text{T})}$ ratios ranging from 0.28121 to 0.28140, with $\epsilon\text{Hf}_{(\text{T})}$ values of +7.8 to +3.6 calculated on the U-Pb zircon age of 2600 Ma. Depleted mantle model (T_{DM}) ages range from 2878-2620 Ma (T_{DM1}) and two-stage model (T_{DM2}) ages from 2779-2611 Ma.

5.1.2 Calc-alkaline granitoids

Sample PAV-VES-1a (14°06'27"N; 77°17'19"E)-Porphyritic monzogranite

Sample PAV-VES-1a is a coarse-grained porphyritic monzogranite collected from a large quarry in the north-eastern outskirts of Pavagada town (Fig. 2). The zircon CL images display euhedral to subhedral morphology with concentric zoning locally overgrown by unzoned zircon. They have low to high U (154-1821 ppm) and Th (75-1042 ppm) contents, with Th/U

467 ratios ranging from 0.20-1.26, indicative of magmatic crystallization. A set of 24 analyses
468 define an upper intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2546 ± 8 Ma (MSWD=0.93) that corresponds
469 to the crystallization age of the porphyritic monzogranite (Fig. 8.2a).

470 *In-situ* Hf isotope analysis of zircons from sample PAV-VES1a record $^{176}\text{Hf}/^{177}\text{Hf}_{(\text{T})}$ ratios
471 ranging from 0.28101 to 0.28116, with corresponding $\epsilon\text{Hf}_{(\text{T})}$ values ranging from +0.5 to -6.1
472 calculated at the U-Pb age of 2546 Ma. Depleted mantle (T_{DM}) model ages range from 3100-
473 2847 Ma (T_{DM1}) and 3032-3433 Ma (T_{DM2}).

474 *Sample KO-18 (14°00'35"N; 77°44'56"E)-Dark grey quartz monzonite*

475 Sample KO-18 was collected from an abandoned quarry ~3 km north of Kodikonda (Fig. 2).
476 It is a coarse-grained dark grey monzogranite with phenocrysts of alkali feldspar. Zircons
477 show prismatic and pyramidal forms with concentric zoning typical of growth from a magma.
478 Evidence for recrystallization can be observed locally (e.g. section 3.1). The U and Th
479 contents and Th/U ratios are range between 166-368 ppm, 105-337 ppm and 0.34 – 0.80,
480 respectively. Uranium–Pb data are discordant and 4 analyses define an upper intercept
481 $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2526 ± 13 Ma (MSWD=0.43, Fig. 8.2b).

482 Zircons from this quartz monzonite record $^{176}\text{Hf}/^{177}\text{Hf}$ ratios ranging from 0.28119 to 0.28135
483 and $\epsilon\text{Hf}_{(\text{T})}$ values ranging from +1.1 to +1.8 at the corresponding age of 2526 Ma. The
484 depleted mantle (T_{DM}) model ages are in the range of 2882-2693 (T_{DM1}) and 3092-2760 Ma
485 (T_{DM2}).

486 *Sample KO-19a (14°09'17"N; 77°45'53"E)-Dark grey quartz monzonite*

487 This sample was collected from an abandoned quarry ~2 km west Puttaparthi (Fig. 2). It is a
488 coarse-grained, dark grey rock that is affected by shearing. Zircon grains have a doubly-
489 terminated prismatic morphology with narrow concentric zones. They record U and Th
490 contents and Th/U ratios in the range of 101-774 ppm, 79-307 ppm, and 0.16 – 0.90,
491 respectively. A set of euhedral zircons (16/19 analyses) define a discordia with an upper
492 intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2573 ± 8 Ma (MSWD=2.4, Fig. 8.2c) and this is interpreted as the
493 time of crystallization.

494 The Hf isotope analyses of the zircon grains record $^{176}\text{Hf}/^{177}\text{Hf}$ ratios ranging from 0.28123 to
495 0.28133 and $\epsilon\text{Hf}_{(\text{T})}$ values ranging from +3.4 to +6.9 at the corresponding age of 2573 Ma.

496 Depleted mantle (T_{DM}) model ages are in the range of 2747-2610 Ma for (T_{DM1}) and 2855-
497 2637 Ma (T_{DM2}).

498 *Sample KO 15 (13°58'08"N; 77°50'35"E)-Porphyritic monzogranite*

499 Porphyritic monzogranite was collected 1 km south of Reddicheruvukatta village along the
500 Kadiri – Kodikonda road (Fig. 2). The sample is a coarse-grained porphyritic monzogranite
501 with cream coloured K-feldspar phenocrysts. Zircon grains are euhedral with oscillatory
502 zoning. They record low to moderate U (76-556 ppm) and Th (49-969 ppm) contents with
503 Th/U ratios of 0.20-1.74, indicating a magmatic origin. Twelve zircons were analysed, and
504 the results are discordant (Fig. 8.2d) defining an upper intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2568 ± 11
505 Ma (MSWD = 0.46), considered to record the time of crystallization.

506 *Sample PAV-VES-1d (14°06'27"N; 77°17'19"E)-Fragmented synplutonic mafic dyke*

507 This sample of synplutonic mafic dyke was collected from the same quarry as sample PAV-
508 VES-1a (see Fig. 2). The zircons are squat to elongated grains and show oscillatory as well
509 as convolute zoning in CL. They have variable U (75-810 ppm) and Th (42-1386 ppm)
510 concentrations, with Th/U values ranging from 0.06 – 2.19 indicative of a magmatic origin.
511 The results (17/21 grains) define an upper intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2535 ± 6 Ma
512 (MSWD = 1.14; Fig. 8.2e).

513 The Hf isotope analyses of zircons from sample PAV-VES1d record $^{176}\text{Hf}/^{177}\text{Hf}_{(T)}$ ratios in
514 the range of 0.28111 to 0.28125, with $\epsilon\text{Hf}_{(T)}$ values ranging from +5.0 to -1.6 at the
515 corresponding age of 2535 Ma. They indicate depleted mantle (T_{DM1}) model ages of 3121-
516 2661- Ma and (T_{DM2}) values of 3139-2737 Ma.

517 *Sample KO-22 (14°04'40"N; 77°17'45"E)-Pink anatectic granite*

518 An anatectic pink granite sample was collected from a quarry ~3 km south of Pavagada along
519 the eastern boundary of the Closepet batholith (Fig. 2). Zircon grains are euhedral to
520 subhedral and are commonly squat with low aspect ratios and partially developed terminal
521 pyramids and pinacoids. Evidence for partial recrystallization includes overgrowths and
522 recrystallized domains in some grains. The grains contain variable contents of U (139-1438
523 ppm) and Th (21-754 ppm) that result in highly variable Th/U ratios 0.12-2.26. Twenty one
524 zircon analyses display two discordant patterns and define two sets of ages, with the older
525 $^{207}\text{Pb}/^{206}\text{Pb}$ data defining an upper intercept age of 3328 ± 14 Ma (MSWD=2.6) and the
526 younger $^{207}\text{Pb}/^{206}\text{Pb}$ age group defining an upper intercept of 2531 ± 6 Ma with MSWD=0.93

527 (Fig. 8.2f). The younger age is interpreted as the time of crystallization of the anatectic melt,
528 whereas the older $^{207}\text{Pb}/^{206}\text{Pb}$ age is interpreted to define inherited zircon from the basement
529 incorporated in the melt.

530 *In-situ* Hf isotope analyses of the zircon grains record $^{176}\text{Hf}/^{177}\text{Hf}_{(\text{T})}$ ratios ranging from
531 0.28066 to 0.28110 and $\epsilon\text{Hf}_{(\text{T})}$ values ranging from -2.0 to -10.7 at the corresponding age of
532 2531 Ma and +1.4 to +4.4 for the older grains at 3328 Ma. The calculated depleted mantle
533 (T_{DM}) ages range from 4064-3158 Ma.

534 Zircon Hf isotope data reveal $^{176}\text{Hf}/^{177}\text{Hf}$ ratios ranging 0.28124 – 0.28133 with $\epsilon\text{Hf}_{(\text{T})}$ values
535 of +1.7 to +6.2. The depleted mantle (T_{DM}) model ages range from 2955-2685 Ma ($T_{\text{DM}2}$).

536 **5.2 Eastern block**

537 *5.2.1 Transitional TTGs and diatexites*

538 *Sample KO-8 (13°43'21"N; 78°20'06"E)-Banded tonalitic gneiss*

539 Sample KO-8 is a dark grey banded tonalitic gneiss collected north of Horsley Hill (Fig.2).
540 The zircon CL images display cores with concentric growth zoning, typical of magmatic
541 growth in a magma, surrounded by recrystallized rims. Overgrowths of dark-CL zircon with
542 euhedral to subhedral forms are commonly present. Most of the analyzed sites contain
543 variable U contents (230-903 ppm, with a few outliers having 16-18 ppm; #9-1 and #9-2) and
544 low Th contents (1-269 ppm). The Th/U ratios range from 0.06 to 0.41. The results define a
545 discordia, with the upper intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2564 ± 13 Ma (Fig.8.3a).

546 *In-situ* Hf isotope data for the zircon reveal $^{176}\text{Hf}/^{177}\text{Hf}$ ratios ranging from 0.28110-0.28120
547 with $\epsilon\text{Hf}_{(\text{T})}$ values of +0.8 to -3.2. The depleted mantle (T_{DM}) model ages range from 2925-
548 2836 ($T_{\text{DM}1}$) and 3240-3015 Ma ($T_{\text{DM}2}$).

549 *Sample KO-7 (13°39'34"N; 78°25'16"E)-Diatexitic gneiss*

550 This sample is a light pinkish-grey diatexitic gneiss collected at the second turning on the
551 way to the top of Horsley Hill (Fig. 2). The CL images of the analysed zircons show dark
552 luminescence with concentric prismatic to pyramidal zones typical of magmatic growth.
553 They record low to moderate U (82-733 ppm) and Th (60-850 ppm) contents, with Th/U
554 ratios ranging from 0.31 to 2.19. Nine grains define a discordia with an upper intercept
555 $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2586 ± 33 Ma (MSWD=0.70), (Fig. 8.3b). This age corresponds to time of
556 magmatic crystallization.

557 Hf isotope data reveals $^{176}\text{Hf}/^{177}\text{Hf}_{(\text{T})}$ ratios ranging from 0.28127 to 0.28132, with $\epsilon\text{Hf}_{(\text{T})}$
558 values from +3.3 to +7.4. The depleted mantle (T_{DM}) model ages range from 2887-2614 Ma
559 (T_{DM2}).

560 *5.2.2 Calc-alkaline granitoids*

561 *Sample KO-10 (13°46'58"N; 78°21'53"E)-Quartz monzodiorite*

562 Dark grey quartz monzodiorite was collected in the north-eastern outskirts of Devalacheruvu
563 (Fig.2). Zircons reveal elongated prismatic morphology with partially developed terminal
564 pyramids. The CL images display concentric oscillatory zoning of magmatic origin. The
565 zircons contain low to moderate U (99-520 ppm) and Th (76-614 ppm) concentrations, with
566 Th/U ratios ranging from 0.80 to 1.52. Results of 15 analyses define a discordant array with
567 an upper intercept $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2566 ± 6 Ma (MSWD=1.5, Fig.8.3c).

568 *In-situ* Hf isotope analyses of zircon reveal $^{176}\text{Hf}/^{177}\text{Hf}_{(\text{T})}$ ratios ranging from 0.28112-
569 0.28123 with $\epsilon\text{Hf}_{(\text{T})}$ values of +2.2 to -1.9. The depleted mantle (T_{DM2}) model ages range
570 from 3166 to 2916 Ma.

571 *Sample KO 29a (13°52'34"N; 78°06'30"E)-Quartz monzodiorite*

572 This sample of quartz monzodiorite was collected from exposures at about 1.5 NW of Marlapalli
573 village (Fig.2). Zircons are magmatic grains with well-developed pyramids and oscillatory
574 zoning; apatite and feldspar inclusions are common. Twenty analyzed zircon grains define a
575 concordia upper intercept age of 2566 ± 8 Ma (Fig.8.3d), which corresponds to the time of
576 magmatic crystallization.

577 *In-situ* Hf isotope analyses of zircon record $^{176}\text{Hf}/^{177}\text{Hf}_{(\text{T})}$ ratios ranging from 0.28115 to
578 0.28129, with $\epsilon\text{Hf}_{(\text{T})}$ values of +3.8 to -1.8. The depleted mantle (T_{DM2}) model ages range
579 from 3160 to 2818 Ma.

580

581 **6. Whole-rock geochemistry**

582 *6.1 Major and trace elements*

583 Analytical methods for major and trace elements are presented in the Appendix. The major and
584 trace element data are presented in Supplementary table 3.

585 A total of 73 samples representing banded dark grey gneisses (transitional TTGs), granitoids
586 (intrusive and anatectic granites), MMEs/synplutonic dykes from the studied corridor of both the
587 Central and Eastern blocks were analysed for major and trace elements. Published data for ca.
588 3300-3000 Ma migmatitic gneisses (Moyen et al., 2003a; Jayananda et al., 2000, 2018) are used
589 for comparison. The migmatitic TTGs, transitional TTGs, sanukitoids and anatectic granites
590 exhibit distinct major and trace element characteristics.

591 The ca. 3360-3000 Ma old migmatitic TTGs are mainly restricted to the Central block and are
592 characterized by high SiO₂ (70-76 wt%) compared to the ca. 2670-2600 Ma dark grey banded
593 gneisses (SiO₂=63-71 wt%) and ca. 2570-2520 Ma granitoids (SiO₂=63-69 wt%). In this study,
594 our use of elemental data from the migmatitic TTGs is restricted to discuss secular changes in
595 petrogenetic and geodynamic processes through time. The transitional TTGs of the Central block
596 contain lower SiO₂ (66 to 71 wt%) and higher MgO (0.97-2.49 wt%) than the transitional TTGs
597 of the Eastern block, which record SiO₂ of 63-71 wt% and MgO of 0.45-2.32 wt%. Further
598 elemental characteristics of the ca. 2570-2520 Ma granitoid plutons of the Eastern Dharwar
599 Craton with sanukitoid affinity have been documented in several studies (Balakrishnan and
600 Rajamani, 1987; Jayananda et al., 1995, 2000, 2018; Moyen et al., 2003a; Sarvothaman, 2001;
601 Dey et al. 2012, 2014, 2016). Consequently, we describe the granitoids with lower SiO₂, higher
602 MgO and incompatible elements from the studied corridor as sanukitoids.

603 On the Ab-An-Or (Fig. 9a) triangular diagram (O'Conner, 1965 with fields by Barker, 1979) the
604 transitional TTGs plot in the tonalite field, with a few samples extending into the granodiorite
605 domain, whereas the >3000 Ma TTGs plot mainly in the trondhjemite field (shown as shaded
606 area in Fig. 9a). The sanukitoids plot in the granodiorite field and extend into the quartz
607 monzonite field, whereas the anatectic granites plot mainly in the granite field and the
608 synplutonic dykes plot in the tonalite-granodiorite and quartz monzonite fields (Fig. 9a). On the
609 K-Na-Ca triangular plot (Fig. 9b), the transitional TTGs define a trend transitional between
610 trondhjemite (TTGs) and calc-alkaline (granitoids), whilst sanukitoids, anatectic granites and
611 syn-plutonic mafic dykes follow the calc-alkaline differentiation trend. The
612 Al₂O₃/(CaO+Na₂O+K₂O) mol versus K₂O/Na₂O mol plot (Fig. 9c), as well as on the
613 Al₂O₃/(FeO_t+MgO) versus CaO/(Na₂O+K₂O) plot (Fig. 9d), shows the transitional TTGs extend
614 from the TTG into the sanukitoid field, whereas the younger sanukitoid and anatectic granites
615 plot in the biotite-granite field. On Harker binary diagrams (Fig. 10), the transitional TTGs,
616 sanukitoids, and anatectic granites define moderate to strong linear trends. The fragmented
617 synplutonic mafic dykes form two groups with one group (hybrid facies?) defining a trend

618 together with the sanukitoids, whereas others form a cluster outside the main trend (Fig. 10). All
619 the studied rocks collectively define linear trends on Harker major element binary plots,
620 implying a genetic link.

621 Among the trace elements, Sr, Ni, Cr, and V of the individual groups (transitional TTGs,
622 sanukitoids, anatectic granite) define moderate to strong negative trends against SiO₂, except for
623 Rb that shows a positive trend. The synplutonic mafic dykes form clusters with a few samples
624 (hybrid facies?) defining a trend together with the sanukitoids. Zirconium displays a very weak
625 negative trend, whereas Ba, Nb and Y show a large spread.

626 Leedy chondrite (Masuda et al., 1973) data were divided by 1.2 with values for Pr, Tb, Ho
627 and Tm interpolated after Taylor and Gordon (1977). The studied rocks from the Central and
628 Eastern blocks show large variations in total rare earth elements (REE) contents and display
629 distinct chondrite normalized REE patterns. The >3000 Ma old migmatitic TTGs in the
630 Central block are characterized by low to moderate total REE (9.2-163.0 ppm) with moderate
631 to highly fractionated REE patterns without any significant Eu anomalies (cf. Jayananda et
632 al., 2018). On the primitive mantle (Sun and McDonough, 1989) normalized multi-element
633 diagram these rocks display weak negative or no Ba-Sr anomalies, but have strong negative
634 Nb-Ta-Ti anomalies that suggest retention of minor plagioclase and also significant titanite
635 and/or ilmenite in the source residue (Jayananda et al., 2018). The transitional TTGs of the
636 Central block display moderate to high total REE (95-325 ppm) with moderate to highly
637 fractionated REE patterns [(La/Yb)_N = 14.26-57.68] and weak negative to positive (or no) Eu
638 anomalies (Eu/Eu* = 0.86-1.10, Fig. 11a). On the primitive mantle normalized multi-element
639 spider diagram the transitional TTGs exhibit positive Sr anomalies (except one sample with a
640 negative anomaly) but without Ba anomalies and with strong negative Nb-Ta-Ti anomalies,
641 implying clinopyroxene, hornblende, titanite and/or ilmenite, with or without garnet and
642 plagioclase, in the source residue (Fig. 11b).

643
644 Data obtained in this study, together with published data (Jayananda et al., 1995, 2000) for
645 sanukitoids from the Central block, reveal high to very high total REE (200-1076 ppm),
646 except for a few samples that show slightly lower total REE (<200 ppm) concentrations.
647 They define poorly to highly fractionated REE patterns, with (La/Yb)_N values ranging from
648 7.51- 45.22 (Fig. 11c) and with minor negative or no Eu anomalies (Eu/Eu* = 0.44-1.08).
649 Further, on the primitive mantle normalized multi-element diagram (Fig. 11d) they show
650 strong negative Nb-Ta-Ti anomalies accompanied by positive Sr, but mostly without Ba

1 651 anomalies (only one sample shows a negative anomaly) and with weakly negative or no Y
2 652 anomalies, suggesting residual clinopyroxene, titanite and/or ilmenite, with variable garnet,
3 653 but without any significant plagioclase in the source residue. The anatectic granites record
4 654 low to high total REE (43-384 ppm; this study and [Jayananda et al., 2000](#)) and show
5 655 moderate to highly fractionated $[(La/Yb)_N = 7.4-46.5]$ patterns with strong negative Eu
6 656 anomalies ($Eu/Eu^* = 0.43-0.52$, [Fig. 11e](#)), together with negative Ba-Sr and Nb-Ta-Ti
7 657 anomalies ([Fig. 11f](#)), indicating major residual plagioclase, biotite, titanite and/or ilmenite,
8 658 but with minor or no residual garnet in the source. The synplutonic mafic dykes in the Central
9 659 block form two distinct patterns of REE. Group I has slightly higher SiO_2 (50-52 wt%) and is
10 660 characterized by low to moderate total REE (32-171 ppm) and flat REE patterns with
11 661 $(La/Yb)_N$ ratios ranging from 1.23 -2.13 ([Fig. 11g](#)). Group 2 has lower SiO_2 (48-59 wt%) and
12 662 high total REE (108-775 ppm) contents, with moderate to strongly fractionated $[(La/Yb)_N$
13 663 $=11.0-37.78]$ patterns ([Fig. 11i](#)). Both groups of synplutonic mafic dykes record either
14 664 negative or positive Ba-Sr anomalies with strongly negative Nb-Ta-Ti anomalies, suggesting
15 665 retention of olivine, clinopyroxene, phlogopite, titanite and /or ilmenite in the mantle source
16 666 ([Fig. 11h & j](#)).

17 667
18 668 In the Eastern block, the transitional TTGs exhibit moderate to high total REEs (72-393 ppm)
19 669 and moderate to strongly fractionated REE patterns $[(La/Yb)_N = 10.76-50.45]$ without any
20 670 significant Eu anomalies ($Eu/Eu^* = 0.84-1.02$, [Fig. 12a](#)). On the primitive mantle normalized
21 671 spider diagram ([Fig. 12b](#)), Sr shows both positive and negative anomalies, with strong
22 672 negative Ba-Nb-Ta-Ti anomalies and weakly negative or no Y anomalies, suggesting variable
23 673 residual plagioclase, clinopyroxene, hornblende, garnet, titanite and/or ilmenite in the source
24 674 residue.

25 675
26 676 The sanukitoids have low to moderate total REE (68-314 ppm) contents but highly fractionated
27 677 $[(La/Yb)_N = 20.87-60.61]$ patterns without any significant Eu anomalies ($Eu/Eu^* = 0.78-1.09$, [Fig.](#)
28 678 [12c](#)). They display positive Ba-Sr anomalies, but strong negative Nb-Ta-Ti anomalies and weak
29 679 positive Y anomalies ([Fig. 12d](#)), suggesting residual clinopyroxene, hornblende, titanite and/or
30 680 ilmenite without any significant residual plagioclase or garnet. The anatectic granites have low to
31 681 high total REEs (26-395 ppm) and show moderate to strongly fractionated REE patterns
32 682 $[(La/Yb)_N = 14.65-66.45]$, with negative Eu anomalies Eu ($Eu/Eu^* = 0.38-0.83$, [Fig. 12e](#)),
33 683 implying residual plagioclase in the source. The anatectic granites display negative Ba-Sr, Nb-
34 684 Ta-Ti anomalies on the primitive mantle normalized multi-element spider diagram ([Fig. 12f](#)),

685 suggesting residual plagioclase, clinopyroxene, biotite, titanite and/or ilmenite. The synplutonic
686 mafic dykes display high total REEs (104-204 ppm) with moderate to highly fractionated REE
687 patterns [(La/Yb)_N = 6.27-25.39] without any significant Eu anomalies (Fig. 12g). The observed
688 positive Sr and Ba anomalies are coupled with negative Nb-Ta-Ti anomalies (Fig. 12h).

689 *6.2 Zircon in-situ Hf isotopes and whole-rock Nd isotopes*

690 Fourteen samples whose zircons were analysed for Lu-Hf isotope compositions reveal significant
691 variability in Hf isotope compositions, with initial ¹⁷⁶Hf/¹⁷⁷Hf_(T) ratios (0.28066 to 0.28135) and
692 εHf_(T) values of +7.8 to -10.7 (Supplementary table 2). The migmatitic gneisses, together with
693 published data (Bidyananda et al., 2016; Dey et al., 2016), have variable initial ¹⁷⁶Hf/¹⁷⁷Hf
694 isotope compositions (0.281236 to 0.280680) with positive εHf_(T) = +3 to +6, indicating slight Hf
695 isotope heterogeneity on the grain scale. The transitional TTGs have initial ¹⁷⁶Hf/¹⁷⁷Hf ratios
696 ranging from 0.28110 to 0.28132 with εHf_(T) = +7.8 to -3.2, indicating significant Hf isotope
697 heterogeneity and composite sources involving depleted to chondritic mantle reservoirs with
698 minor crustal contamination, particularly in the western part of the corridor. Sanukitoids show
699 significant variation in initial ¹⁷⁶Hf/¹⁷⁷Hf compositions (0.28101 – 0.28135), with εHf_(T) = +6.9
700 to -6.1, which could be related to Hf isotope heterogeneity involving depleted to enriched mantle
701 with ancient crustal contamination. The anatectic granite from the Central block shows evolved
702 initial ¹⁷⁶Hf/¹⁷⁷Hf_(T) composition (0.28066), with εHf_(T) = -2.0 to -10.7, indicating reworking of
703 ancient crust. The initial ¹⁷⁶Hf/¹⁷⁷Hf_(T) versus zircon age plot (Fig. 13a) and εHf_(T) versus zircon
704 ages (Fig. 13b) reveal a chondritic to depleted mantle source for the magmatic precursors of the
705 ca. 3360-3000 Ma migmatitic gneisses, with traces of significant ancient crustal input (in the
706 Central block), depleted to chondritic mantle for the transitional TTGs, a chondritic to enriched
707 mantle source, with or without crustal contamination, for the sanukitoids and ancient to short
708 lived crust for the anatectic granites.

709 Whole rock Nd isotope data are presented in Supplementary table 4. The studied gneisses and
710 granitoids display significant variation in Nd isotope compositions.

711 In the Central block, the 3271 Ma migmatitic basement gneisses record εNd_(T=3271Ma) values of
712 +1.97 to -0.5, with Nd_{TDM} ages of 3400-3335, which are close to the U-Pb zircon age, suggesting
713 derivation of the magmatic precursors from short-lived mafic crust. The ca. 2670-2600 Ma
714 transitional TTG have εNd_(T) ranging from +2.75 to +1.71, with corresponding Nd_{TDM} ages of
715 2919-2735 Ma, which are slightly older than the U-Pb zircon ages. These values indicate the

1 716 origin of the rocks either from short-lived mafic crust or from composite sources involving the
2 717 mantle as well as short-lived mafic crust. The four samples of sanukitoids show $\epsilon\text{Nd}_{(T=2545 \text{ Ma})}$
3 718 values ranging from +1.99 to -1.26, with Nd_{TDM} ages of 2700-3000 Ma. The fragmented
4 719 synplutonic mafic dyke has a slightly negative $\epsilon\text{Nd}_{(T=2535 \text{ Ma})}$ value of -0.82 with an older Nd_{TDM}
5 720 age of 2920 Ma compared to the zircon age of 2535 Ma. The ϵNd values of the sanukitoids and
6 721 associated synplutonic dykes imply their derivation from mantle reservoirs contaminated by pre-
7 722 existing crust. The anatectic pink granite, together with published data, has high negative
8 723 $\epsilon\text{Nd}_{(T=2535)}$ values of -5.44 to 0.0, with corresponding model ages of 3400-2700 Ma (Moyen,
9 724 2000; Dey et al., 2014; Jayananda et al., 2000, 2018), suggesting its origin through reworking of
10 725 ancient as well as newly-formed juvenile felsic crust. Five samples of transitional TTG in the
11 726 Eastern block record positive $\epsilon\text{Nd}_{(T)}$ values ranging from +2.6 to +0.53, with Nd_{TDM} ages ranging
12 727 from 2706-2798 Ma, which are slightly older than the zircon crystallization ages. These values
13 728 suggest a short-lived juvenile source or depleted to chondritic mantle derived melts contaminated
14 729 by lower crust. The sanukitoids record positive $\epsilon\text{Nd}_{(T)}$ ranging from +1.21 to +0.93 with
15 730 corresponding Nd_{TDM} ages of 2878-2798 Ma, indicating their derivation from a juvenile source
16 731 without much ancient crustal contamination. In summary, whole-rock Nd isotope data indicate
17 732 major mantle input, with minor ancient crustal signatures during Neoproterozoic accretion of the
18 733 Central block, whereas such ancient crustal signatures are not evident in the Eastern block (Fig.
19 734 13c).

20 735 **7. Discussion**

21 736 *7.1 Effects of secondary processes on elemental compositions*

22 737 The studied corridor was affected by shear deformation, fluid flow and amphibolite to granulite
23 738 facies metamorphism (Chardon et al., 2008, 2011; Jayananda et al., 2006, 2013a). Therefore, it
24 739 is necessary to evaluate the effects of fluid flow and metamorphism on element mobility prior to
25 740 using the elemental data to constrain petrogenetic processes, including the composition of
26 741 potential source regions as well as the geodynamic context of magma generation.

27 742 The modification of elemental composition of some samples by secondary processes including
28 743 fluid induced hydrothermal alteration and/or medium- to high-grade metamorphism, has resulted
29 744 in significant variation in K_2O and Na_2O contents, the spread of values for the large-ion
30 745 lithophile elements (LILE), including Rb, Ba and to some extent Sr, on the Harker diagrams, and
31 746 variations revealed on the multi-element spider diagrams. However, relatively immobile

747 elements like REE, Nb, Ta, Y, Zr and Hf display smooth patterns on primitive mantle
748 normalized multi-element diagrams, therefore these elements can be used as robust tracers to
749 constrain potential sources of the different gneisses and granitoids of the studied corridor.
750 Further, [Polat and Hoffmann \(2003\)](#) proposed that rocks with high loss of ignition (LOI >6.0
751 wt%) and large Ce anomalies (positive or negative) may be significantly affected by secondary
752 alteration processes. However, most of the analysed samples in the present study exhibit much
753 lower LOI (<1.0 wt%) and do not show any Ce anomalies. This precludes significant alteration
754 of elemental composition of the analysed samples by post-magmatic secondary processes.
755 Further, [Jayananda et al \(2000, 2015, 2018\)](#) addressed the issue of the effect of secondary
756 processes on the elemental and isotopic compositions and concluded that only LILE (Cs, Rb, K,
757 Ba and Sr) are disturbed to some extent, but that REE and high field strength elements (Nb, Ta,
758 Ti, Zr, Hf, Sc, Y) were not affected. Consequently, primary geochemical signatures of most of
759 the elements have not been disturbed and these elements can be used to discuss petrogenesis and
760 geodynamic context of magma genesis.

761 *7.2 Petrogenesis and sources*

762 The zircon U-Pb data presented in this study, together with published data, show that migmatitic
763 TTGs, transitional TTGs, sanukitoids, and synplutonic mafic dykes and anatectic granites
764 correspond to crust building events at ca. 3360-3000 Ma, 2670-2600 Ma, 2573-2526 Ma and
765 2530 Ma, respectively.

766 *7.2.1 Ca. 3360-3000 Ma Migmatitic TTGs*

767 The origin of TTGs is a topic of much debate, with diverse models proposed in the past two
768 decades ([Smithies et al., 2000](#); [Rapp et al., 2003](#); [Martin et al., 2005, 2014](#); [Moyen, 2011](#); [Polat, 2012](#);
769 [Hoffmann et al., 2011](#); [Jayananda et al., 2015, 2018](#)). The origin of Archean TTGs from
770 different cratons has been explained by hydrous melting of down-going oceanic crust ([Martin, 1986](#);
771 [Feng and Kerrich, 1992](#); [Foley et al., 2002](#)), melting of subducted oceanic plateaus ([Martin et al., 2014](#)),
772 or melting of thickened oceanic arc crust ([Hoffmann et al., 2011](#); [Polat, 2012](#);
773 [Jayananda et al., 2015, 2018](#)). It is generally accepted that TTGs are generated by hydrous
774 melting of mafic rocks at different depths ([Moyen, 2011](#)). However, the composition of the
775 mafic crust and tectonic context of TTG magma genesis remains strongly debated ([Johnson et al., 2019](#)).
776 The >3000 Ma old migmatitic TTGs in the Central block of the Dharwar Craton are
777 characterized by high SiO₂ (>70 wt%, except for one sample containing 66.0 wt%), high
778 Na₂O/K₂O (>3.0), and variable but moderate to high LREE, Ba, and Sr, together with low Mg#,

779 Ni and Cr. These characteristics preclude melting of a down-going oceanic slab in a subduction
780 zone as the source, because “oceanic crust formed at a spreading centre is strongly depleted and
781 cannot account for the high content of incompatible elements” (Smithies et al., 2000; Hoffmann
782 et al., 2011; Jayananda et al., 2015). Alternatively, two-stage melting involving melting of
783 subducted oceanic crust to generate new oceanic arc crust, followed by hydrous melting of the
784 subsequently thickened arc crust at different depths, may be a possibility (Hoffmann et al., 2011;
785 Polat, 2012). Partial melting of thickened oceanic arc crust at different depths with variable
786 hornblende, clinopyroxene, plagioclase, garnet, titanite/ilmenite in the residue could account for
787 the observed high REE contents, low Mg# (<0.30), moderate to high (La/Yb)_N and Sr/Y, variable
788 Ba-Sr and negative Nb-Ti anomalies on spider diagrams (Jayananda et al., 2018). Thus, high
789 incompatible element contents, together with highly fractionated REE, negative Nb-Ta-Ti
790 anomalies of the >3000 Ma migmatitic TTGs in the Central block probably originated by
791 melting of short-lived thickened mafic oceanic arc crust at different depths. This agrees with the
792 Nd isotope data that record a Nd_{TDM} age of 3400-3365 Ma, whereas U-Pb zircon defines a
793 slightly younger age of 3360-3271 Ma, with εNd_(T) values of +1.97 to -0.5, suggesting a two-
794 stage melting process for the origin of the TTGs.

795 7.2.2 Ca. 2670-2600 Ma Transitional TTGs

796 The lower SiO₂ (68.0 wt%), higher MgO (1.3 wt%), CaO (3.06 wt%) and K₂O (2.40 wt%), and
797 lower Na₂O/K₂O ratios (1.92) but higher Mg# (>0.39) of the transitional TTGs compared to the
798 ca. 3360-3000 Ma migmatitic TTGs from the Central block [average SiO₂ (72.42 wt%), MgO (
799 0.58 wt%), CaO (2.15 wt%), K₂O (1.68 wt%), reveal their transitional character between TTGs
800 and calc-alkaline granitoids of sanukitoid affinity, as do their higher Na₂O/K₂O ratios (>3.0), but
801 lower Mg# (<0.30)], together with their abundant biotite, hornblende and local clinopyroxene
802 (Jayananda et al., 2018). The origin of Neoproterozoic transitional TTGs has been explained by
803 melting of mafic lower crust (Champion and Smithies, 2003b; Jayananda et al., 2006; Dey et al.,
804 2016) or by melting of composite sources involving both arc crust and sub-arc mantle
805 (Jayananda et al., 2018). Moderate to strong fractionation trends of Fe₂O₃, MgO, MnO, CaO,
806 TiO₂, and P₂O₅ versus SiO₂ on Harker diagrams (see Fig. 10) indicate removal of clinopyroxene,
807 hornblende, minor plagioclase, and ilmenite/titanite as fractionating phases or their retention in
808 the source residue.

809 Hydrous melting of a down-going oceanic slab in a subduction zone can be considered for the
810 origin of transitional TTGs. However, their high content of incompatible elements (K, Rb, Ba,

811 Sr, Zr, Th, U) precludes slab melting in a subduction zone, as oceanic crust generated at
812 spreading centres is highly depleted in incompatible elements. Melting of such depleted source
813 regions generates melts with much lower incompatible element contents (Smithies et al., 2009;
814 Nagel et al., 2012). An alternate possibility is a two-stage melting process involving melting of
815 hydrated oceanic crust in a subduction zone producing oceanic arc crust, followed by melting of
816 such thickened arc crust to generate magmatic precursors of the transitional TTGs. However,
817 recent petrological and experimental data show that melting of thickened mafic arc crust at
818 different depths generates TTG magmas (Hoffmann et al., 2011; Nagel et al., 2012; Polat, 2012)
819 with high SiO₂, and low K₂O, Rb, Ba, Sr, MgO, Ni and Cr contents. On the other hand,
820 transitional TTGs contain low SiO₂ but conversely high MgO, K₂O, Rb, Ba, Sr, Ni, and Cr
821 contents, coupled with abundant hornblende ± clinopyroxene, calcic plagioclase, and K-feldspar,
822 which cannot be explained by melting of thickened arc crust alone. This feature is reflected on
823 the 3*CaO-Al₂O₃/FeO_t+MgO-5*(K₂O/Na₂O) triangular plot (Laurent et al., 2014), wherein the
824 transitional TTGs plot both in the low-K as well as the high-K mafic source fields (Fig. 14a),
825 implying the involvement of both thickened arc crust as well as an enriched source, possibly
826 from the mantle. Involvement of both mafic arc crust as well as a mantle source (depleted to
827 chondritic) is also reflected in the Sr+Ba versus FeO_t+MgO binary plot (Laurent et al 2014)
828 where the transitional TTGs plot in the crustal source field and extend into the field of enriched
829 mantle (Fig. 14b). Whole rock Nd isotopic data ($\epsilon\text{Nd}_{(T)} = 0.52$ to 4.23) suggest composite
830 sources involving both short-lived mafic arc crust as well as a mantle (chondritic to depleted)
831 source. This is also in agreement with *in-situ* Hf isotope data of zircons [$\epsilon\text{Hf}_{(T)} = -0.67$ to 5.61].
832 The strongly fractionated REE patterns (14.2-57.6, Fig. 11a) together with the Sr/Y versus Y
833 (Fig. 15a) and (La/Yb)_N versus Yb (Fig. 15b) plots suggest variable amounts of clinopyroxene,
834 hornblende and garnet (0-30 %) in the source residue. Furthermore, the strong negative Nb and
835 Ti anomalies in the multi-element spider diagrams indicate retention of ilmenite and/or titanite in
836 the source residue, whereas positive Sr anomalies, or the absence of anomalies, imply minor or
837 no significant residual plagioclase. Although the transitional TTGs of the Central and Eastern
838 blocks share broadly similar elemental characteristics, they do show significant variation in some
839 elements. The higher contents of Ba-Sr, MgO, Ni, Cr and V of rocks from the Central block
840 imply a dominantly mantle input (chondritic to slightly depleted), whilst slightly lower contents
841 in the Eastern block suggest lesser mantle input and dominantly arc crust involvement.

842 It is important to constrain the melting conditions and mineralogy of the source residue. The
843 strong negative Nb, Ta and Ti anomalies on the multi-element spider diagrams (see Figs. 11, 12)

844 suggest residual titanite and/or ilmenite. The Nb/Ta, Zr/Hf, and Zr/Sm ratios can be used as
845 powerful tracers to characterize source compositions and residual mineralogy, as these elemental
846 pairs have different partition coefficients for residual phases like amphibole, clinopyroxene,
847 titanite, ilmenite and rutile (Foley et al., 2002; Rapp et al., 2003; Xiong et al., 2007; Zhou et al.,
848 2014). Amphibole has a major control on Nb/Ta, Zr/Sm and Zr/Hf ratios, as well as heavy (Yb)
849 to middle (Gd, Dy) REE. Consequently, melts generated with hornblende in the source residue
850 would be expected to have low Nb/Ta, Gd/Yb, and Dy/Yb, but high Zr/Sm ratios (Zhou et al.,
851 2014). On the contrary, garnet retained in the source residue increases the Sr/Y, La/Yb and
852 Zr/Sm ratios (Davidson et al., 2007). The Nb/Ta *versus* Zr/Sm plot suggests dominant amphibole
853 and clinopyroxene, with variable amounts of garnet and ilmenite, in the source residue (Fig. 16).
854 In addition, melts with residual garnet are generally expected to show a negative correlation of
855 Al₂O₃ with SiO₂, but the very weak correlation of Al₂O₃ with SiO₂ (see Fig. 10) precludes major
856 residual garnet in the source, which is also in agreement with the absence of a strong positive
857 correlation between Sr/Y versus (La/Yb)_N (Figure not presented). In contrast, the plots of
858 (La/Yb)_N versus (Yb)_N and Sr/Y versus Y suggest variable (0-30 %) residual garnet in the source
859 (see Figs. 15a,b). The low Dy/Yb (1.99-3.22), Gd/Yb (2.22 -5.38), and Nb/Ta (5.24-14.03)
860 ratios, together with variable Zr/Sm values (11.88-66.40), indicate amphibole and/or
861 clinopyroxene together with variable garnet in the melt residue (Jenner et al., 1994; Klemme et
862 al., 2002). This is in agreement with the observed low to high Cr/V values (0.49 -2.04),
863 suggesting major amphibole and/or clinopyroxene, plus variable (0-30 %) garnet in the residue.
864 The absence of significant Eu anomalies with only minor negative or rare positive Eu/Eu*
865 values, implies magma generation close to or below the plagioclase stability field (Fig. 11a, c).

866 In summary, the elemental and Hf-Nd isotope data indicate that the magmatic protoliths of the
867 transitional TTGs were derived by melting of composite sources involving juvenile thickened
868 mafic arc crust and underlying sub-arc mantle, with clinopyroxene, hornblende, garnet (0 -30%),
869 minor or no plagioclase, and titanite/ilmenite in the source residue.

7.2.3 *Ca. 2570-2535 Ma Sanukitoids*

871 Sanukitoids are intermediate to felsic magmatic rocks enriched in both incompatible (LILE,
872 HFSE, REE) and compatible (Mg, Ni, Cr) elements which post-date the accretion of TTGs.
873 Numerous studies have investigated sanukitoids from different cratons and have shown
874 geochemical and isotope signatures corresponding to major juvenile additions, with or without
875 minor ancient crustal input (for reviews see Shirey and Hanson, 1984; Jayananda et al., 1995,

876 2000, 2018; Smithies and Champion, 2000; Moyen et al., 2003a; Martin et al., 2005, 2010;
1 877 Lobach-Zhuchenko et al., 2008; Halla et al., 2005; Wang et al., 2009; Heilimo et al., 2010;
2 878 Laurent et al., 2014).

6 879 The data presented here, together with published data (Jayananda et al., 1995, 2000; Moyen,
7 880 2000; Moyen et al., 2003; Dey et al., 2014, 2016) from the Central and Eastern Dharwar Craton,
8 881 reveal a wide range of compositions (SiO_2 , 50-71 wt%), suggesting that their parental magmas
9 882 evolved through differentiation processes. The negative trends of MgO , Fe_2O_3 , MnO , CaO , TiO_2
10 883 and P_2O_5 against SiO_2 suggest olivine, clinopyroxene, apatite and ilmenite were fractionating or
11 884 residual phases. The strong negative trends of Ni, Cr and V against SiO_2 also reflect the possible
12 885 role of olivine and clinopyroxene, whereas negative Sr indicates plagioclase fractionation. The
13 886 REE patterns, without significant Eu anomalies, suggest simultaneous crystallization of
14 887 clinopyroxene/hornblende and plagioclase. The depletion of Nb-Ta-Ti might be related to the
15 888 involvement of titanite/ilmenite. The dominantly positive Sr anomalies on the multi-element
16 889 spider diagrams, together with the absence or weakly negative Y anomalies, indicate the variable
17 890 but dominant role of clinopyroxene/hornblende, with or without minor residual garnet.

29 891 The low SiO_2 (<65 wt%) contents, coupled with high Mg-number (>0.39), Ni (20-46 ppm) and
30 892 Cr (110-134 ppm) contents, preclude an ancient crustal source and suggest mafic to ultramafic
31 893 rocks as the potential source. These characteristics, together with high to very high contents of
32 894 incompatible elements (K, Rb, Ba, Sr, Zr, Nb, Y, REE) even at low SiO_2 (<55 wt%), argue for
33 895 enriched ultramafic sources. The high SiO_2 (>64 wt%) and low-Mg# (<0.39) of transitional
34 896 TTGs means they are inconsistent as a potential source, as reflected in the 3^*CaO -
35 897 $\text{Al}_2\text{O}_3/(\text{FeO}_t+\text{MgO})-5^*(\text{K}_2\text{O}/\text{Na}_2\text{O})$ ternary (Fig. 14a) diagram (Laurent et al., 2014). Interaction
36 898 of slab melts generated at different depths with the overlying mantle wedge during their ascent is
37 899 considered to be the source of sanukitoid magmas (Martin et al., 2005, 2010). Melting of mafic
38 900 rocks at higher pressures leaving garnet-rich amphibolite or eclogite with rutile residue
39 901 (Klemmen et al., 2002; Rapp et al., 1991, 2003) would generate melts with high Al_2O_3 (>15
40 902 wt%) but low Mg# and transition and HFSE. This is contrary to the observed low Al_2O_3 (13-15
41 903 wt%), but conversely high Mg# (0.39-0.61) and transition elements, thus precluding high
42 904 pressure melting of mafic rocks, leaving garnet-rich amphibolite or eclogite residue.
43 905 Furthermore, several studies have shown that melting of thickened oceanic arc crust at different
44 906 depths would generate melts of low- to high-Al TTGs as revealed by elemental data and
45 907 experimental results with low Mg# (<0.40) but variable Sr/Y and $(\text{La}/\text{Yb})_N$ ratios (Smithies et
46 908 al., 2009; Hoffmann et al., 2011; Nagel et al., 2012; Polat, 2012; Jayananda et al., 2015, 2018). =

909 Such a scenario fails to account for the observed low SiO₂ (<60 wt%), but high Ni, Cr, REE,
910 Ba+Sr, and HFS elements (Zr, Hf, Nb, Ta, Sc, Y). Therefore, melting of mafic rocks with
911 eclogite or garnet-amphibolite residue and thickened oceanic arc crust cannot account for the
912 observed low SiO₂, and high contents of both compatible and incompatible elements.
913 Consequently, involvement of an enriched peridotitic mantle source is required for the
914 generation of sanukitoid melts. Major enriched mantle input is reflected in the Sr+Ba versus
915 FeO_T+MgO diagram, where the majority of the samples plot in the enriched mantle field, with a
916 few samples extending into the crustal melting field (Fig. 14b), which probably corresponds to
917 minor crustal input or contamination.

918 The high Th/Ta, Th/Nb, La/Nb, Ba/Nb, and Ba/La ratios of the sanukitoids suggest possible
919 enrichment of sub-arc mantle reservoirs (Shirey and Hanson, 1984; Smithies and Champion,
920 2000). *In situ* Hf [$\epsilon\text{Hf}_{(T)} = +4.5$ to -3.2] isotope data for zircons, together with whole rock Nd
921 isotope data [$\epsilon\text{Nd}_{(T)} = +1.93$ to -1.26] for the sanukitoids, are compatible with mantle (enriched
922 or not) reservoirs with or without minor crustal input. Traces of ancient crustal input is evident in
923 the sanukitoids in the Central block, as revealed by angular migmatitic enclaves and negative
924 $\epsilon\text{Hf}_{(T)}$ and $\epsilon\text{Nd}_{(T)}$ values of a few samples, whereas sanukitoids in the Eastern block show
925 positive $\epsilon\text{Hf}_{(T)}$ and $\epsilon\text{Nd}_{(T)}$ signatures, which is agreement with the absence of ancient (>3000
926 Ma) crust. The high to very high content of incompatible elements (LILE, LREE, HFSE), even
927 at the lowest SiO₂ (50 wt%), is also consistent with their derivation from enriched mantle
928 reservoirs.

929 Elemental and isotope data reveal major mantle input in the origin of the sanukitoid magmas, but
930 it is necessary to characterize the depth of melting and residual mineralogy of the mantle source.
931 The moderately fractionated REE patterns of the majority of samples, without or with weak
932 negative or positive Eu anomalies, together with positive Sr anomalies and negative Nb-Ta-Ti
933 anomalies without any significant Y anomalies (see Figs. 11c, d), suggest melting close to or
934 below the plagioclase stability field, with residual olivine, clinopyroxene,
935 amphibole/phlogopite?, titanite and/or ilmenite, but minor residual garnet. This is consistent with
936 the Sr/Y versus Y and (La/Yb) versus Yb plots (Figs. 15a, b) suggesting minor or no residual
937 garnet. The Nb/Ta versus Zr/Sm plot indicates low pressure melting with possible co-existing
938 residual clinopyroxene, amphibole, and ilmenite (Figure not presented). To summarize, the field,
939 elemental and isotope data of the sanukitoids from the Central block indicate derivation from
940 chondritic to enriched mantle reservoirs that interacted with ancient crust, whereas the

941 sanukitoids from the Eastern block were derived from a less enriched or chondritic to depleted
942 mantle source that probably interacted with young juvenile crust, as reflected by the field and
943 isotopic data.

944 7.2.4 Ca. 2535 Ma Synplutonic mafic dykes

945 Syn-plutonic mafic dykes correspond to the terminal phase of juvenile magmatism in the Central
946 and Eastern Dharwar Craton and are synchronous with emplacement of anatectic granite, as
947 revealed by field evidence (see Fig. 3j) and U-Pb zircon ages (see Fig. 8.2e). Most of these
948 dykes are fragmented which indicates interaction with the granitoid host, as revealed in the field
949 by the presence of biotite-rich rims. Petrographic characteristics (K-feldspar and quartz
950 inclusions in clinopyroxene/hornblende) suggest these mafic melts were injected into
951 crystallizing magma chambers at different stages of crystallization of the host (Jayananda et al.,
952 2009, 2014; Gireesh et al., 2012). Two groups are defined on Harker diagrams (Fig. 10), wherein
953 higher SiO₂ (55-60 wt%) samples (hybrid facies?) define differentiation trends along with the
954 sanukitoids, whereas the low SiO₂ (48-51 wt%) samples form a cluster that corresponds to mafic
955 injections without much interaction with the host, as they were probably injected during an
956 advanced crystallization stage (>70% crystals) of the host. Their low SiO₂ contents, together
957 with moderate, high to very high content of compatible (MgO, Ni, Cr) as well as incompatible
958 (K, Ba, Sr, REE, Th, U, Zr, Y, Nb) elements, cannot be attributed to either melting of a down-
959 going slab in a subduction zone or to the melting of oceanic arc crust, but imply involvement of a
960 peridotitic mantle source. Their high contents of incompatible elements cannot be explained by
961 crustal contamination, as the samples with the lowest SiO₂ (<50 wt%) contain the highest
962 incompatible elements compared to the surrounding granitoid host. On the 3*CaO-
963 Al₂O₃/(FeO_T+MgO)-5*(K₂O/Na₂O) ternary plot (Fig. 14a, after Laurent et al., 2014) these rocks
964 plot in the high-K mafic source region, suggesting enriched reservoirs that were also reflected on
965 the Sr+Ba versus FeO+MgO diagram (Fig. 14b; after Laurent et al., 2014) where they plot as
966 vaugnerites, corresponding to enriched lithospheric mantle reservoirs, with two samples
967 extending into the deep seated OIB field. *In situ* zircon Hf [$\epsilon\text{Hf}_{(T)}$ = -1.6 to +5.0] isotope and
968 whole-rock Nd [$\epsilon\text{Nd}_{(T)}$ = +0.88 to -0.7] indicate mainly enriched mantle reservoirs, except for
969 three samples that show depleted mantle signatures (lower K, Ba, Sr, LREE). Three samples
970 from the Central block with low to moderate total REE (32-171 ppm) show flat REE patterns
971 without Eu anomalies, suggesting their derivation from shallow depleted to chondritic mantle.
972 Their relatively unfractionated to highly fractionated REE patterns [(La/Yb)_N = 1.22 to 37.78],

973 together with low to moderate Sr/Y values and strong negative Nb-Ta anomalies, argue for
974 melting of peridotitic mantle at different depths (shallow to deep) with variable co-existing
975 residual olivine, clinopyroxene, phlogopite, and ilmenite/rutile, without any significant garnet in
976 the residue. The large variation in elemental abundances (low to very high K, Ba, Sr, and LREE)
977 in the syn-plutonic mafic dykes of the Central block can be attributed to heterogeneous sub-arc
978 mantle, whereas moderate K, Ba, Sr, and LREE in the Eastern block could be related to
979 moderately enriched homogeneous arc mantle.

980 In summary, the 2670-2600 Ma transitional TTGs and 2570-2520 Ma sanukitoid intrusions are
981 juvenile additions that are spatially and temporally related, as reflected in their elemental and
982 isotopic compositions. However, a progressive increase of K₂O/Na₂O, Rb, Ba, Sr, Nb, Zr, Ti, P
983 and Y from transitional TTGs to sanukitoids indicates their petrogenetic link, particularly the
984 evolution of their mantle source reservoirs.

985 *7.2.5 Ca. 2535-2520 Ma Anatectic granites*

986 These rocks are present at the periphery of large juvenile granitoid intrusions and are found as
987 small dykes traversing the migmatitic TTGs, as well as the banded transitional TTGs. Field
988 evidence, such as diffuse contacts with the surrounding gneisses, suggests their derivation by
989 remelting of crustal rocks. Zircon U-Pb ages show that anatectic granite emplacement
990 immediately preceded emplacement of the hot sanukitoid magmas. Field evidence, such as
991 evidence of mingling between anatectic granite and sanukitoid (Jayananda et al., 2014), and fluid
992 inclusion data (Santosh et al., 1991), show that juvenile calc-alkaline magmas provided heat and
993 a fluid source for deep crustal melting. The moderate to strong negative trends of TiO₂, Fe₂O₃,
994 MgO, CaO, Na₂O, P₂O₅, Sr, Ni, Cr, and V on the Harker diagrams was likely related to
995 involvement of plagioclase, hornblende/clinopyroxene, titanite/ilmenite and apatite as
996 fractionating or residual phases. Their high SiO₂ (>70 wt%), K, Rb, Ba, and LREE contents, but
997 low Sr, could be attributed to anatexis of crustal rocks as suggested by the 3*CaO-
998 Al₂O₃/(FeO_T+MgO)-5*(K₂O/Na₂O) triangular diagram (see Fig. 14a). Isotopic data presented in
999 this study, together with published data [$\epsilon\text{Hf}_{(T)} = -4.3$ to -15.0) and $\epsilon\text{Nd}_{(T)} = -0.5$ to -7.0],
1000 indicate their derivation by melting of heterogeneous ancient crust (>3000 Ma) as well as newly
1001 formed juvenile crust (ca. 2700-2600 Ma). The moderate to strongly fractionated REE patterns
1002 with strong negative Eu (Eu/Eu* = 0.34-0.81) anomalies, together with negative Ba, Sr, Nb-Ta,
1003 and Ti anomalies, suggest major residual plagioclase + hornblende \pm clinopyroxene \pm garnet,
1004 with titanite/ilmenite. On the contrary, the large variation in (La/Yb)_N values (7.03-20, excluding

1005 two outliers with values >50) and Sr/Y (8-82), together with the absence of Y anomalies,
1006 preclude significant garnet in the source residue. Experimental work and geochemical modelling
1007 suggest that melting of alumina-rich crustal source rocks can generate higher (La/Yb)_N and Sr/Y
1008 ratios (Moyen, 2011), which may account for the few high (La/Yb)_N and Sr/Y values in the
1009 present study. In summary, the anatectic granites formed by reworking of ancient (>3000 Ma)
1010 together with newly formed crust (2700-2600 Ma) and are spatially associated with the
1011 emplacement of the hot sanukitoid magmas.

1012 *7.3 Multi-stage crust accretion, continental growth and reworking*

1013 Global zircon U-Pb ages, together with *in situ* Hf isotope compositions and combined with
1014 whole-rock Nd isotope data, suggest high rates of continental crust production during the
1015 Archean through differentiation of the mantle (Belousova et al., 2010; Condie and Aster, 2010;
1016 Condie et al., 2017; Dhuime et al., 2012; Roberts and Spencer, 2014). Furthermore, zircon age
1017 peaks have been attributed to the generation of large continental masses through collision of
1018 crustal blocks produced at subduction zones (Hawkesworth et al., 2010; Condie, 2014). Global
1019 ‘super’ events of continental crust production have been documented at ca. 2700 Ma, 1900 Ma
1020 and 1200 Ma (Condie, 2000).

1021 The zircon U-Pb ages presented in this study, together with published zircon ages from the
1022 Central and Eastern Dharwar Craton, show episodic juvenile crustal accretion and multi-stage
1023 continental growth at ca. 3360-3200 Ma, 3000-2960 Ma, 2700-2600 Ma and 2570-2520 Ma
1024 (Friend and Nutman, 1991; Peucat et al., 1993, 2013; Balakrishnan et al., 1999; Rogers et al.,
1025 2007; Chardon et al., 2011; Bidyananda et al., 2016; Anand and Balakrishnan, 2010; Anand et
1026 al., 2014; Dey et al., 2014, 2016, 2018; Jayananda et al., 1995, 2000, 2013a, 2018; Khanna et
1027 al., 2014, 2016; Ratheesh Kumar et al., 2016; Mohan et al., 2013, 2014; Nasheeth et al., 2015).
1028 These juvenile accretion events are spatially associated with crustal reworking, including
1029 metamorphism at ca. 3200 Ma, 3000 Ma, 2620 Ma, 2520 Ma (Mahabaleswar et al., 1995; Peucat
1030 et al., 2013; Jayananda et al., 2011, 2013a). Furthermore, U-Pb zircon data from Neoproterozoic
1031 juvenile crust indicate a crude age zonation from west to east, wherein the crust progressively
1032 become younger with more primitive juvenile signatures and lesser degrees of enrichment to the
1033 east (see Fig. 17).

1034 The oldest preserved Archean basement is confined to the westernmost part of the Central block
1035 represented by the migmatitic TTGs that accreted in two stages, corresponding to significant
1036 continental growth. Our SHRIMP U-Pb zircon data and published zircon age data for the

1037 migmatitic gneisses from the northern part of the study area (Chardon et al., 2011; Jayananda et al., 2013a; Nasheeth et al., 2015), suggest that the oldest crust corresponds to the earliest continental growth at ca. 3380-3230 Ma. Furthermore, detrital zircons from metapelite from the western (Jayananda et al., 2013a) and eastern margins of the Closepet batholith record U-Pb ages ranging from 3350-3200 Ma (Jayananda unpub data). In the southern part of the Central block, U-Pb zircon ages for the migmatitic gneisses from the Bangalore-Kolar and Kabbaldurga – B.R. Hills region reveal two stages of magmatism that contributed to significant crustal growth at ca. 3360-3200 Ma and 3000-2960 Ma (Friend and Nutman, 1991; Jayananda et al., 2000; Peucat et al., 2013; Mahabaleswar et al., 1995; Bidyananda et al., 2011, 2016; Ratheesh Kumar et al., 2016). A few younger U-Pb ages documented from rims on old zircons at ca. 2700 Ma, ca. 2600 Ma and ca. 2500 Ma, coincide with ca. 2700 Ma greenstone volcanism, 2600 Ma transitional TTG accretion and 2540 Ma sanukitoid magmatism, respectively (Balakrishnan et al., 1999; Jayananda et al., 2000, 2013a; Mojzsis et al., 2003; Yang and Santosh, 2015; Peucat et al., 2013). *In situ* Hf isotope [$\epsilon\text{Hf}_{(T)} = +5$ to -7)] data for zircon and whole-rock Nd isotope data [$\epsilon\text{Nd}_{(T)} = +3$ to -1.6] data suggest significant continental growth at ca. 3360-3150 Ma and 3000-2960 Ma involving juvenile addition as well as minor ancient crustal contamination (Bhaskar Rao et al., 2003; Peucat et al., 2013; Ratheesh Kumar et al., 2016; Bidyananda et al., 2016). In summary, the U-Pb/Pb-Pb ages of magmatic and detrital zircons from the oldest basement in the Central block show two stages of continental growth at ca. 3360-3150 Ma and 3000-2960 Ma, prior to Neoproterozoic juvenile accretion. However, the 3380-3150 Ma komatiite-dominated greenstone assemblages with contemporaneous TTGs documented in the Western block (Jayananda et al., 2008; Maya et al., 2017) are absent in the Central block. This suggests an independent crustal history for the Central and Western blocks of the Dharwar Craton. Although some workers (e.g. Bidyananda et al., 2011, 2016) argued that both the Western and Eastern Dharwar Craton (based on their data presented only for the Central block up to the Kolar greenstone belt) evolved simultaneously but that fundamental differences exist in lithological assemblages, thermal records, age zonation and accretion histories of the three micro-blocks (Jayananda et al., 2018). The zircon U-Pb ages when combined with *in situ* Hf and Nd isotope data for transitional TTGs, together with published data, reveal a major episode of juvenile felsic crust addition contributing to high rates of continental growth in both the Central and Eastern blocks at ca. 2670-2580 Ma (Krogstad et al., 1991; Balakrishnan et al., 1999; Mojzsis et al., 2003; Chardon et al., 2011; Dey et al., 2014, 2016; Yang and Santosh, 2015). U-Pb zircon data indicate widespread transitional TTG accretion in the Central block at ca. 2670-2600 Ma, but at ca. 2590-2580 Ma in the Eastern

1070 block (see Figs. 8.1 and 8.3). This indicates that transitional TTG accretion in the Eastern block
1071 was initiated during the waning stages of transitional TTG formation in the Central block. On the
1072 other hand, published geochronological data for adjoining greenstone volcanic assemblages (see
1073 Fig. 17) reveal bimodal age distributions with ca.2745- 2670 Ma mafic to felsic volcanism and
1074 minor felsic volcanism at 2570-2545 Ma (Balakrishnan et al., 1990; Zachariah et al., 1995;
1075 Nutman et al., 1996; Balakrishnan et al., 1999; Naqvi et al., 2002; Anand and Balakrishnan,
1076 2010; Sarma et al., 2012; Jayananda et al., 2013b; Khanna et al., 2014, 2016; Dey et al., 2018;
1077 Yang and Santosh, 2015). This indicates that transitional TTG accretion was immediately
1078 preceded by ca. 2745-2670 Ma greenstone volcanism in the Central and Eastern blocks.
1079 Consequently, ca. 2745-2670 Ma greenstone volcanism and 2670-2600 Ma transitional TTG
1080 accretion together contributed to large-scale continental growth.

1081 Calc-alkaline juvenile granitoids of sanukitoid affinity cover nearly 40 percent of the surface
1082 area of the Central and Eastern blocks. The U-Pb zircon ages of the present study, together with
1083 published ages on calc-alkaline granitoids, define major outbursts of crust accretion that
1084 contributed to peak continental growth between 2570-2520 Ma (Peucat et al., 1993, 2013;
1085 Jayananda et al., 1995, 2000, 2013a; Chardon et al., 2002, 2011; Rogers et al., 2007; Anand et
1086 al., 2014; Mohan et al., 2014; Dey et al., 2012, 2014, 2016; Yang and Santosh, 2015). The Nd-
1087 Sr-Hf isotope data of these granitoids suggest major differentiation of heterogeneous mantle
1088 reservoirs with variable, but minor, older crustal input. This juvenile magmatic accretion was
1089 contemporaneous with felsic volcanism in the adjoining greenstone belts (Jayananda et al.,
1090 2013b; Dey et al., 2015; Yang and Santosh, 2015). Consequently, calc-alkaline magmatism of
1091 sanukitoid affinity, together with contemporaneous felsic volcanism, have likewise contributed
1092 to peak continental growth between 2570-2520 Ma (see Fig. 17).

1093 Combined geochronological and isotope data reveal a complex evolutionary history with multi-
1094 stage crustal growth involving juvenile magmatic accretion, crustal reworking and
1095 metamorphism. Isotopic age data record multiple stages of crustal growth (ca. 3360-3150 Ma,
1096 3000-2960 Ma, 2700-2600 Ma, 2570-2520 Ma) in the Central block, whilst only two episodes of
1097 crustal growth are evident in the Eastern block at 2700-2670 and 2580-2535 Ma. These major
1098 episodes of crustal growth were spatially linked to crustal reworking and medium- to high-grade
1099 metamorphism, as revealed by texturally controlled *in situ* monazite and U-Pb zircon ages. In the
1100 Central block, the earliest thermal event is dated at ~3200-3140 Ma, which corresponds to the
1101 terminal phase of migmatitic TTG accretion (this study; Jayananda et al., 2013a). The second
1102 thermal event, defined by upper amphibolite to granulite facies metamorphism, occurred at ca.

1103 2960 Ma as revealed by discordant zircon ages (Mahabaleswar et al., 1995; Jayananda and
1104 Peucat, 1996) and this is spatially linked to 3000 Ma crustal accretion. Metamorphism and
1105 dehydration melting of pelitic rocks under granulite facies to UHT conditions (800-900°C, 5-7
1106 kbar), documented in the Central block at 2620 Ma (Jayananda et al., 2011, 2013a), coincide
1107 with the accretion of transitional TTG. However, the pre-2500 Ma thermal event is not
1108 documented in the Eastern block. Both the Central and Eastern blocks underwent a coeval
1109 amphibolite to granulite facies thermal event at ca. 2513 Ma, followed by slow cooling to 2400
1110 Ma (Jayananda et al., 2013a; Peucat et al., 1989, 2013). In summary, the Eastern Dharwar Craton
1111 preserves records of multi-stage crustal growth and reworking events.

1112 Globally, 2700 Ma zircon age peaks have been attributed to high rates of continental growth as a
1113 result of craton collisions (Condie et al., 2017). However, such zircon age peaks have not been
1114 documented in the Dharwar Craton. Only a few whole-rock isochrons and zircon ages of ca.
1115 2700 Ma have been reported for the greenstone volcanics (Balakrishnan et al., 1990, 1999;
1116 Khanna et al., 2014, 2016; Naqvi et al., 2002; Jayananda et al., 2013b). In contrast, the U-Pb
1117 zircon ages of 2670-2600 Ma for the transitional TTGs, and the following 2570-2520 Ma
1118 outburst of calc-alkaline granitoid magmatism, correspond to major juvenile crustal addition that
1119 contributed to large-scale Neoproterozoic continental growth. Consequently, the major global peak
1120 in continental growth at ca. 2700 Ma is craton specific and not evident in the Dharwar Craton.

1121 *7.4 Mantle evolution through the Neoproterozoic: Implications for enrichment of source reservoirs*

1122 Globally, several studies have attributed mantle enrichment to (i) melting of sediments in
1123 subduction zones; (ii) interaction of slab-derived melts with overlying mantle (Prouteau et al.,
1124 2001; Sajona et al., 2000; Moyen et al., 2003a; Wang et al., 2009); (iii) fluid flux from slab
1125 dehydration (Heilimo et al., 2010); or (iv) metasomatism related to plume/asthenosphere
1126 upwelling (Oliveira et al., 2011; Jayananda et al., 2018). Magmas generated from sub-arc mantle
1127 that was modified by melting of subducted sediments are characterized by low Sr/Ce, Ce/Yb and
1128 Sr/Y ratios (Tatsumi, 2006; Shimoda et al., 1998), whereas melts generated from arc mantle
1129 enriched by fluids derived from subduction zones exhibit Sr/Ce (~20) and Ce/Pb values <20
1130 (Chauvel et al., 1995). The wide range of Sr/Ce ratios of 2.74-16.96, Ce/Pb values of 3.43-9.76
1131 and Sr/Y values ranging from 30-160 obtained in this study are not fully consistent with mantle
1132 enrichment either through fluids or sediment recycling in a subduction zone. This is supported by
1133 positive $\epsilon\text{Nd}_{(T)}$ and $\epsilon\text{Hf}_{(T)}$ values for the majority of the sanukitoid samples. Slab-derived fluids
1134 or melt induced mantle enrichment have been proposed by many workers (Kamber et al., 2002;

1135 [Moyen et al., 2003](#); [Komiya and Maruyama, 2007](#)), wherein enrichment of supra-subduction
1 1136 zone mantle results in depletion of Ta and Hf relative to La and Sm by fluid induced
2 1137 metasomatism.
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6 1138 In the Eastern Dharwar Craton, the differentiation of crust from the mantle in the Neoproterozoic
7 1139 occurred at ca. 2750 -2520 Ma in the form of greenstone volcanism, transitional TTGs, calc-
8 1140 alkaline magmatism of sanukitoid affinity and synplutonic mafic dykes. It is therefore important
9 1141 to constrain the coupled evolution of the crust–mantle system and enrichment of mantle
10 1142 reservoirs through time.
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16 1143 *7.4.1 Constraints from greenstone volcanism (2746-2670 Ma)*
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19 1144 Greenstone volcanism is bimodal in age distribution, with the ultramafic and mafic volcanism
20 1145 and associated minor felsic volcanism, representing the earlier episode at ca. 2750-2670 Ma,
21 1146 whereas dominantly felsic volcanism occurred between 2580-2545 Ma. Elemental and isotope
22 1147 (Pb, Sr, Nd, Hf) data for the greenstone volcanics provide important constraints on mantle
23 1148 evolution, particularly the enrichment of source reservoirs to the transitional TTGs, sanukitoids
24 1149 and syn-plutonic mafic dykes. The Central block recorded only the earlier episode of volcanism
25 1150 (komatiite-basalt-andesite-dacite) in the Sandur-Ramagiri-Penakacherla-Kustagi-Hungund
26 1151 greenstone belts at ca. 2750-2670 Ma ([Zacharaiah et al., 1995](#); [Nutman et al., 1996](#); [Balakrishnan](#)
27 1152 [et al., 1999](#); [Naqvi et al., 2002](#)). On the contrary, two episodes of volcanism are documented in
28 1153 the Eastern block, with ca. 2700 -2696 Ma komatiite-basalt volcanism in the Kolar, Kadiri,
29 1154 Veligallu, Jonnagiri, Gadwal and Hutti greenstone belts, and minor ca. 2580-2545 Ma felsic
30 1155 volcanism in the Kolar, Kadiri, Gadwal, Hutti and Peddavura greenstone belts.
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42 1156 The 2746-2702 Ma komatiitic to high-Mg basaltic assemblages in the Central block appear to be
43 1157 generated in hot-spot environments, forming oceanic islands/plateaus, whereas the associated
44 1158 calc-alkaline basalts, adakites–rhyolites originated by melting of young subducted oceanic crust
45 1159 in a convergent margin setting ([Zacharaiah et al., 1996a](#); [Manikyamba and Kerrich, 1992](#); [Naqvi](#)
46 1160 [et al., 2002](#)). Elemental ratios of the Sandur greenstone komatiites and basalts ($Nb/Th <8$ to >8)
47 1161 have been interpreted in terms of their derivation from a heterogeneous plume source, with the
48 1162 higher contents of incompatible elements of associated adakites and dacites attributed to a
49 1163 hybridized mantle wedge source in an arc setting, which, upon eruption, accreted to a plume-
50 1164 derived oceanic plateau ([Manikyamba et al., 2008](#)). The volcanic-dominated ca. 2746-2702 Ma
51 1165 Ramagiri-Penakacherla-Kushtagi-Hungund greenstone sequences are heterogeneous. The
52 1166 alkaline (~8.0 wt%) character of basalts from the Penakacherla greenstone belt, together with
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1167 their high Mg# (0.70-0.58), TiO₂ (2.3-2.1 wt%), and REE contents, when coupled with their
1168 fractionated patterns [(La/Yb)_N = 23-29] and weak negative Nb anomalies and higher Zr/Hf
1169 ratios (Kerrick and Manikyamba, 2012), all point to a primitive mantle source. Immediately
1170 south of this belt, variable LREE contents and Nd isotope data ($\epsilon\text{Nd}_{(T)} = +2.0$ to $+3.5$), with
1171 some anomalous values linked to post-crystallization alteration) of metabasalts from the
1172 Ramagiri greenstone belt (Fig. 2) are attributed to distinct mantle sources (depleted to
1173 enriched?), with oceanic slab-derived fluids playing a major role in enrichment of LREE and
1174 LILE (Zacharaiah et al., 1997). To summarize, the elemental and isotope data of the 2750-2700
1175 Ma volcanic sequences in the Central block reveal mantle heterogeneity (depleted to enriched),
1176 with enrichment of mantle reservoirs through slab-derived fluids at ca. 2700 Ma.

1177 In the Eastern block, incompatible trace element contents and Nd and Pb isotope compositions of
1178 komatiite and high-Mg basalts from the Kolar greenstone belt reveal heterogeneous sources
1179 (LREE depleted in the west and LREE enriched to the east), with long-term depletion of the
1180 mantle and LREE enrichment prior to melting of the mantle to form the eastern volcanics
1181 (Balakrishnan et al., 1990; Krogstad et al., 1989). More recent U-Pb zircon data, together with
1182 Nd isotope data, for the Kolar greenstone volcanics reveal ancient crustal contamination
1183 (Jayananda et al., 2013b). Immediately north in the Kadiri greenstone belt (Fig. 2), two types of
1184 basalts are recognised, one with flat LREE and the other with LREE-enriched patterns, Both
1185 display Th, U, and LILE enrichment with negative Nb, Zr, Hf, and Ti anomalies, together with
1186 high Th/Nb, La/Nb and Ce/Yb ratios, attributed to enrichment of arc mantle through slab-derived
1187 fluids and sediment melting (Manikyamba et al., 2015b). Furthermore, Dey et al. (2016) showed
1188 that volcanic suites in the Kadiri greenstone belt have flat REE patterns with minor negative Nb-
1189 Ta anomalies, interpreted as rocks generated in a mantle plume-derived oceanic plateau, whilst
1190 high Mg# (0.00-0.52), Cr, and Ni, coupled with HFSE depletion and LREE enrichment, was
1191 attributed to their derivation from a metasomatized mantle wedge overlying a subducting slab in
1192 a continental margin setting. Further north, compositionally different (high-Mg, low-Nb, high
1193 Nb) basalts from the Jonnagiri greenstone belt show enrichment in LREE/LILE and depletion in
1194 HFSE, which have been attributed to partial melting of a hybridized mantle wedge through slab
1195 dehydrated fluids (Manikyamba et al., 2014).

1196 Elemental data (high Mg# 0.89, Cr = 2290–3855 ppm and Ni = 604–966 ppm), together with
1197 depletion of REE and $\epsilon\text{Hf}_{(T=2696\text{ Ma})} = +3-0$ to $+6.5$ for ultramafic volcanics from the Veligallu
1198 greenstone belt, reveal long-term depletion of the mantle reservoirs (Khanna et al., 2016).

1199 However, Nd isotopes [$\epsilon\text{Nd}_{(T)} = +0.6$ to $+1.1$], coupled with negative Nb anomalies and
1200 elevated Th/Yb and LREE contents of basalts from this belt suggest derivation from slightly
1201 depleted to metasomatized mantle through subducted slab-derived fluids in an arc setting
1202 (Dey et al., 2018). In the central part of the Eastern block, Hf-Nd isotope compositions Hf
1203 [$\epsilon\text{Hf}_{(T)} = +1.6$ to $+8.7$]; Nd [$\epsilon\text{Nd}_{(T)} = -0.1$ to $+3.0$] of volcanic assemblages (tholeiitic, calc-
1204 alkaline basalts to boninites) from the Gadwal greenstone belt are consistent with a depleted
1205 mantle source that evolved over a protracted period.

1206

1207 In the northern part of the Eastern block, metabasalts from the ca. 2700 Ma Hutti greenstone belt
1208 have both enriched and depleted LREE signatures, together with variable Nb, Ta, P, and Ti
1209 anomalies, high $(\text{La}/\text{Yb})_N$ and Zr/Sm, but low Nb/Ta, that have been explained through
1210 derivation from heterogeneous mantle reservoirs (depleted to chondritic or slightly enriched)
1211 whereby mantle wedge enrichment was caused by slab-derived fluids (Manikyamba et al., 2009).
1212 In addition, whole rock Pb, Sr and Nd isotope data for the ca 2700 Ma basaltic rocks from the
1213 Hutti greenstone belt indicate long term depletion of mantle reservoirs (Anand and Balakrishnan,
1214 2010).

1215 In summary, the incompatible elemental data, together with Pb, Sr, Nd, and Hf isotope data
1216 presented above, argue for the existence of heterogeneous (depleted to chondritic to enriched)
1217 mantle reservoirs at ca. 2700 Ma and likely enrichment of the mantle through slab-derived
1218 fluids/melts or sediment melting in a subduction zone. Elemental and Nd isotope data for the ca.
1219 2700 -2670 Ma greenstone volcanics in the Central block indicate a high degree of enrichment of
1220 the mantle, with possible contribution from detritus derived from >2700 Ma pre-existing crust.
1221 The degree of enrichment of the source was less in the Eastern block, as revealed by elemental
1222 and isotope data from the ca. 2700 and 2580-2545 Ma volcanic assemblages that were erupted
1223 either as oceanic plateaus or in intra-oceanic arc settings.

1224 *7.4.2 Constraints from transitional TTGs (2670-2600 Ma) and sanukitoids (2570-2520 Ma)*

1225 The U-Pb zircon ages presented in this study, together with published data, show that transitional
1226 TTG magmatism at 2670-2600 Ma was immediately followed by 2570-2520 Ma calc-alkaline
1227 magmatism of sanukitoid affinity, including syn-plutonic mafic dykes. A shift in elemental
1228 composition occurred with the change from transitional TTGs to sanukitoids, with an increase in
1229 abundance of both incompatible elements (Rb, K, Ba, Sr, Zr, Y, and LREE) and compatible
1230 elements (Mg#, Ni and Cr). This implies that transitional TTGs and sanukitoids are genetically

1231 linked with an increase in peridotitic mantle input at ca. 2670-2520 Ma. Furthermore, elemental
1 1232 and Nd-Hf isotope data for these two juvenile magmatic suites in the Central block originated
2 1233 from depleted to highly enriched reservoirs, whereas the magmatic suites in the Eastern block
3 1234 were derived from depleted to chondritic reservoirs. When coupled with distinct thermal records
4 1235 and accretion histories, including greenstone volcanism (Manikyamba and Kerrich, 2012), it
5 1236 reveals that large-scale mantle heterogeneity existed in the Eastern Dharwar Craton during the
6 1237 Neoproterozoic. Thus, it is necessary to account for this mantle heterogeneity and the evolution of
7 1238 mantle sources for the transitional TTGs and sanukitoids.

15 1239 In the Eastern Dharwar Craton, strain fabric analysis, petrological data and elemental and isotope
16 1240 constraints favour the convergence of lithosphere in a plate tectonic regime, leading to building
17 1241 of oceanic and continental arcs at ca. 2700 Ma (Balakrishnan et al., 1999; Chadwick et al., 2000,
18 1242 2007; Manikyamba et al., 2017; Dey et al., 2015; Jayananda et al., 2018). During the
19 1243 Neoproterozoic, melting of subducted oceanic crust along with continent-derived detritus in an arc
20 1244 setting would generate melts enriched in incompatible elements. These enriched slab melts could
21 1245 interact and enrich overlying arc mantle in incompatible elements during their ascent. Melting of
22 1246 the down-going subducted slab at deeper levels (~80-100 km) would produce much hotter mafic
23 1247 magmas that, upon their ascent to the crust-mantle boundary, would induce melting of enriched
24 1248 sub-arc mantle and the base of the newly-formed arc crust, generating the magmatic precursors
25 1249 of the transitional TTGs at ca. 2670-2600 Ma. Furthermore, elemental data for the enriched
26 1250 mafic volcanics (e.g. alkaline basalts in the Penakacherla greenstone belt; Kerrich and
27 1251 Manikyamba, 2012) indicate possible derivation from ocean island basalt (OIB) sources and
28 1252 these could also enrich the shallow mantle during their ascent. Consequently, the magmatic
29 1253 precursors of transitional TTGs originated from composite sources enriched by slab-derived
30 1254 magmas and through deep mantle-derived enriched mafic magmas.

45 1255 Thus, transitional TTG accretion was immediately followed by sanukitoid magmatism between
46 1256 2570-2520 Ma in an arc setting and involving a dominantly mantle source, with only minor
47 1257 ancient crustal or newly formed arc crust input. The majority of the studied sanukitoid samples
48 1258 display depletion of Ta relative to La, but do not show depletion of Hf relative to Sm. Thus,
49 1259 mantle enrichment cannot be explained entirely by slab-derived fluids. Instead, melt induced
50 1260 enrichment of a mantle source would lead to depletion of Ta compared to La and enrichment of
51 1261 Hf relative to Sm (LaFleche et al., 1998), as documented for the Taishan sanukitoids from the
52 1262 North China Craton (Wang et al., 2009). The sanukitoid samples from the Eastern Dharwar
53 1263 Craton exhibit $(\text{Ta/La})_N$ values of 0.13-0.89 and $(\text{Hf/Sm})_N$ values of 0.21 -0.98 and they plot in

1264 the field of melt-induced mantle enrichment, with a few samples extending into the field of fluid-
1265 related subduction metasomatism (Fig. 18, after LaFleche et al., 1998). Although it is difficult to
1266 provide a straightforward criterion to discriminate fluid or melt involvement in peridotitic mantle
1267 enrichment, it nevertheless appears that interaction of melts with peridotitic mantle played a
1268 significant role in the study area, as revealed by the $(\text{Ta/La})_N$ versus $(\text{Hf/Sm})_N$ plot (Fig. 18),
1269 although the role of fluids cannot be totally ruled out. In the Eastern Dharwar Craton, continued
1270 subduction between 2700-2600 Ma, with eventual slab breakoff, led to asthenosphere upwelling,
1271 causing widespread melting of enriched sub-arc mantle and thereby generating hot enriched
1272 sanukitoid magmas. Upon their emplacement, these magmas variably interacted with both
1273 ancient crust and newly-formed juvenile crust in the Central block, but only with young juvenile
1274 crust in the Eastern block. Although the sanukitoids in the Central and Eastern blocks show
1275 similar elemental characteristics, those in the Central block show strong enrichment, as well as
1276 extensive interaction with old crust, as also indicated by the field relations (Fig. 3a), and
1277 distinctive elemental features, including several negative $\epsilon\text{Nd}_{(T)}$ and $\epsilon\text{Hf}_{(T)}$ values (see
1278 supplementary tables 2 & 4) when compared with the Eastern block. This is consistent with the
1279 U-Pb zircon ages and isotope data from the Eastern block, which indicate the absence of ancient
1280 (>2.75 Ga) crust. Syn-plutonic mafic dykes were injected during the terminal stage of sanukitoid
1281 magmatism and the majority of samples show moderate to very high contents of K, Rb, Ba, Sr,
1282 LREE, Nb, Zr, Hf, Ta, and Ti despite their low SiO_2 contents (see Figs. 10, 11g, h, j; 14b)
1283 implying the existence of enriched mantle at the end of the Archean. Consequently, the above
1284 features show progressive enrichment of mantle reservoirs throughout the Neoproterozoic, probably
1285 through slab-derived melts/fluids (see Fig. 18).

7.5 Evolving geodynamic processes through the Neoproterozoic: implications for assembly of crustal blocks, orogenic plateau formation and cratonization

1288 The geodynamic context of Archean crustal accretion, continental growth and craton formation
1289 are much debated topics. Numerous models, including stagnant lid through mobile lid to
1290 emergence of modern style plate tectonics, have been proposed for Archean crust formation and
1291 continental growth (for reviews see Bedard, 2018; Wyman, 2018; Cawood et al., 2018 and
1292 references therein). Multidisciplinary studies involving petrological data, strain fabric analysis,
1293 and numerical, thermo-mechanical and analogue modelling have provided important insights
1294 into Archean geodynamic processes as well as the mechanical behaviour of the Archean crust-
1295 mantle system (Smithies et al., 2005; Chardon et al., 2009, 2011; Gerya, 2014; Sizova et al.,
1296 2014, 2015; Thebaud and Rey, 2013; Korenaga, 2018; Moyen and Laurent, 2018).

1297 Geodynamic evolution of Archean crust in the Dharwar Craton has likewise been the subject of
1298 much discussion over the past two decades. Contrasting models based on petrological and
1299 elemental data have been proposed for the Neoproterozoic evolution of the Eastern Dharwar Craton
1300 (Jayananda et al., 2000, 2013, 2018; Harish Kumar et al., 2003; Maibam et al., 2016;
1301 Manikyamba and Kerrich, 2012; Dey et al., 2014, 2018; Santosh and Li, 2018; Pandey et al.,
1302 2018), although only a few models are based on regional structural analysis (Bouhallier et al.,
1303 1993, 1995; Chadwick et al., 2000, 2007; Chardon et al., 1996, 1998, 2002, 2008, 2011) and
1304 geophysical data (Gupta et al., 2003; Borah et al., 2014; Ravi Kumar et al., 2018; Mandal et al.,
1305 2018; Vijaya Rao et al., 2015). The three most popular models include horizontal motion of
1306 plates similar to modern style plate tectonics (Chadwick et al., 2000, 2007; Santosh and Li,
1307 2018), mantle plume-based vertical motions (Bouhallier et al., 1993, 1995; Chardon et al., 1996,
1308 1998, 2002; Jayananda et al., 2000), and a combination of plume-arc processes (Dey et al., 2018;
1309 Harish Kumar et al., 2003; Jayananda et al., 2008, 2013, 2018; Manikyamba and Kerrich, 2012).
1310 Further, based on detailed strain fabric analysis and zircon geochronology, Chardon et al. (2011)
1311 proposed westward convergence of hot orogenic crust that would generate sub-horizontal
1312 lamination and strong seismic reflectivity (Vijay Rao et al., 2015; Mandal et al., 2018).

1313 The studied corridor through the Central and Eastern Dharwar Craton records a complex
1314 geological history and any model proposed has to account for the observed west to east age
1315 zonation patterns, changes in the composition of felsic rocks from TTGs, through transitional
1316 TTGs, to sanukitoids, the low pressure and high temperature metamorphic record and seismic
1317 reflectivity. Horizontal tectonic models, but with different polarity of subduction involving
1318 either westward oblique convergence (e.g. Chadwick et al., 2000, 2007) or eastward subduction
1319 (Bidyananda et al., 2016; Santosh and Li, 2018; Li et al., 2018b), have been proposed to explain
1320 Neoproterozoic crustal accretion and northward-trending fabrics. The westward convergence of
1321 oceanic crust beneath the Western Dharwar Craton is consistent with the observed age zonation
1322 patterns, crustal flow, metamorphic record (low-P and HT to UHT) to the east, but fails to
1323 provide a satisfactory explanation for ca. 2700 Ma komatiite volcanism documented in the
1324 Sandur, Kolar and Veligallu greenstone belts. On the other hand, based on petrologic and
1325 elemental data some workers have proposed eastward convergence of oceanic lithosphere
1326 (Ratheesh Kumar et al., 2016; Santosh and Li, 2018; Li et al., 2018b). A combined plume-arc
1327 accretion model was proposed for the 2700-2500 Ma period of evolution in the Eastern Dharwar,
1328 where komatiite-tholeiite volcanism was attributed to a plume with TTG accretion occurring in
1329 the intervening arcs (Harish Kumar et al., 2003; Manikyamba and Kerrich, 2012; Jayananda et

1330 al., 2013b, 2018; Dey et al., 2018). We discuss the evolving Neoproterozoic geodynamic processes
1331 in terms of the combined plume–arc model, but followed by orogenic plateau formation through
1332 assembly of micro-blocks and final cratonization within the regional framework of a Neoproterozoic
1333 hot orogen (Chardon et al., 2011).

1334 Geological and geophysical data reveal thick old crust (40-54 km) in the Western block,
1335 relatively thin crust (38-33 km) with both old and young juvenile components in the Central
1336 block, and thin (32-35 km) juvenile crust in the Eastern block. This observation is consistent
1337 with the distribution of moderate- to high-pressure metamorphic assemblages (Rollinson et al.,
1338 1981; Iswar Kumar et al., 2016; Jayananda et al., 2013a) and magmatic epidote-bearing ca.3000
1339 Ma plutons in the Western block (Jayananda et al., 2019) and low-pressure, but high to ultra-
1340 high temperature assemblages, in the Central and Eastern blocks. The Chitradurga boundary
1341 shear zone separating the Western and Central blocks is interpreted as a major tectonic boundary
1342 (Chadwick et al., 2000; Chardon et al., 2011; Jayananda et al., 2006, 2013b) where westward
1343 convergence of hot oceanic lithosphere has resulted in addition of hot juvenile magmas, causing
1344 extensive melting and lateral constructional flow of hot orogenic crust on weaker mantle
1345 lithosphere (Chardon et al., 2011), and producing strong seismic reflection patterns (Mandal et
1346 al., 2018). In this scenario, the Chitradurga boundary shear zone acted as a ‘wall’ between the
1347 already cratonized colder Western block and hot crust undergoing lateral constrictional flow on
1348 weaker mantle in the Central block between 2570-2520 Ma (Chardon et al., 2011). The section
1349 of the Central and Eastern blocks that was the focus of this study displays distinct thermal
1350 records and accretionary histories prior to 2600 Ma (Jayananda et al., 2013a). However, both
1351 these crustal blocks formed part of a wide Neoproterozoic hot orogen between 2600-2500 Ma. We
1352 propose a model of secular changes in geodynamic evolution through hotspot magmatism,
1353 leading to oceanic plateaux formation, followed by westward convergence of intervening slabs
1354 producing the arc crust and eventual assembly of crustal blocks into a cratonic framework in the
1355 context of wide hot orogen (Chardon et al., 2011). In this model (Fig. 19a) melting at a mantle
1356 hot spot between ca. 2750-2700 Ma beneath ancient (3300-3000 Ma) crust of the Central block
1357 generated komatiite and alkali basalt magmas, leading to development of the Sandur and
1358 Ramagiri-Penakacherla greenstone basins, whilst komatiite to high-Mg basalt magmas erupted in
1359 oceanic environments producing volcanic plateaux (e.g. the Kolar, Kadiri, Veligallu, Gadwal
1360 greenstone belts) in the Eastern block. Simultaneous westward convergence of intervening
1361 oceanic crust beneath the Western and Central blocks led to the development of the ca. 2740 -
1362 2650 Ma Chitradurga back-arc basin (Kumar et al., 1996; Russel et al., 1991; Nutman et al.,

1363 [1996; Jayananda et al., 2013b](#)) and the 2740 Ma Javagondanahalli basin ([Jayananda et al., 2011b](#))
1364 along the eastern margin of the Western block and the Sandur-Ramagiri-Penakacherla-Hungund
1365 belt in an active continental margin setting ([Fig. 19a](#)). To the east, hotspot volcanism, followed
1366 by westward convergence of intervening oceanic crust beneath the Central block, produced
1367 intermediate to felsic magmas that upon emplacement built arc crust adjacent to mafic to
1368 ultramafic crust in the continental (e.g. Sandur) and oceanic (e.g. Kolar, Veligallu and Gadwal)
1369 regions at ca. 2700 Ma ([Fig. 19b](#)). Melting of down-going slabs at depth, with significant input
1370 of continent-derived detritus, produced fluidal hot mafic magmas. Ascent of these hot mafic
1371 magmas interacted with, and variably-enriched, the sub-arc mantle, as well as causing melting of
1372 both sub-arc mantle and the overlying newly-formed thickened arc crust, leading to generation of
1373 the magmatic precursors to the transitional TTGs between 2670-2600 Ma ([Fig. 19c](#)). These
1374 magmas were emplaced beneath ancient crust in the Central block and newly-formed oceanic arc
1375 crust in the Eastern block ([Fig. 19d](#)). Emplacement of these transitional TTG magmas caused hot
1376 metamorphism of the ancient crust in the Central block close to UHT conditions at 2620 Ma
1377 ([Jayananda et al., 2011a](#)). Continued subduction led to assembly of the Western, Central and
1378 Eastern blocks and eventual slab breakoff led to upwelling of hot asthenosphere, causing
1379 widespread low-degree melting of enriched mantle producing hot sanukitoid magmas between
1380 2570-2520 Ma ([Fig. 19e](#)). These hot magmas were emplaced into already hot ancient and young
1381 juvenile arc crust, causing extensive crustal melting that is consistent with horizontal
1382 contractional flow of hot crust on a hot weaker mantle, with orogenic plateau formation, subdued
1383 topography and strong seismic reflectivity ([Mandal et al., 2018](#)). This is also consistent with
1384 tectonic fabric analysis that shows lateral contractional flow of hot orogenic crust up to the
1385 Chitradurga boundary shear zone between 2565-2500 Ma and 3D mass distribution through
1386 longitudinal flow accommodating the juvenile materials added to the crust during the
1387 Neoproterozoic hot orogen ([Chardon et al., 2011](#)).

1388 The hot juvenile magmas induced melting of both ancient and newly-formed juvenile crust
1389 (transitional TTGs) and generated anatectic magmas ([Jayananda et al., 2000, 2018](#))
1390 corresponding to biotite granites ([Dey et al., 2018; Moyen et al., 2013](#)). Crystallization of
1391 voluminous mantle-derived sanukitoid magmas in deep crustal magma chambers resulted in the
1392 development of shrinkage fractures at mantle depths due to volume loss, which in turn caused
1393 low degree decompression melting of highly-enriched sub-arc mantle, producing low viscosity
1394 enriched mafic magmas that were rapidly injected into crystallizing magma chambers where
1395 their interaction led to mixing and mingling of magmas and formation of MME or fragmented

1396 syn-plutonic mafic dykes depending on the crystallinity state of the host magma (Jayananda et
1397 al., 2009, 2014). Continued heat and fluid flow associated with mantle upwelling also induced
1398 low pressure, high to ultra-high temperature metamorphism and final cratonization of Archean
1399 crust between 2510-2480 Ma (Peucat et al., 1993, 2013; Jayananda et al., 2013a).

8. Conclusions

1400 The integrated field, petrological, zircon U-Pb and *in situ* Lu-Hf, elemental and whole rock
1401 Nd isotope data presented here from a wide crustal corridor in the Central and Eastern
1402 Dharwar Craton define episodic continental growth leading to craton formation, wherein the
1403 gneisses and granitoids form markers recording this transition and fingerprinting a shift in
1404 geodynamic processes through the Archean. The major conclusions arising from this study
1405 are as follows:

1. Zircon U-Pb ages, together with *in situ* Lu-Hf isotopes and whole-rock Sm-Nd isotope data, reveal multi-stage crustal growth in the Central and Eastern blocks of the Dharwar Craton at ca. 3360-3200 Ma, 3000-2960 Ma, 2700-2600 Ma and 2570-2520 Ma.
2. A major outburst of juvenile felsic crustal addition at ca. 2670-2600 Ma is represented by transitional TTG, followed at 2570 -2520 Ma by calc-alkaline magmatism of sanukitoid affinity, which contributed to peak continental growth. The preceding ca. 2750-2700 Ma greenstone volcanism originated from depleted to chondritic mantle reservoirs and contributed to an earlier episode of significant crustal growth.
3. Elemental data, *in situ* Hf isotope, and whole rock Nd isotope data indicate that the magmatic precursors of the transitional TTGs formed by melting of composite sources involving dominantly chondritic mantle and mafic arc crust, whereas the sanukitoids with syn-plutonic mafic dykes originated from chondritic to enriched mantle with or without minor crustal input. Anatectic granites were generated by reworking of ancient and newly formed juvenile crust.
4. The documented west to east age zonation, structural patterns, petrological, elemental and isotope data, together with seismic reflection data, suggest westward convergence of hot oceanic lithosphere leading to assembly of micro-blocks. Continued convergence with eventual slab break-off led to asthenosphere upwelling that induced widespread mantle melting and juvenile magmatic additions, resulting from lateral flow of ductile hot crust on hot weaker mantle and leading to formation of orogenic plateaux with subdued topography, as indicated by strain fabric data and strong seismic reflectivity in the Eastern Dharwar Craton.

- 1434 5. Juvenile calc-alkaline magmatism was spatially linked to extensive crustal reworking,
1 1435 high-temperature/low-pressure metamorphism and final cratonization in the late
2 1436 Neoproterozoic.

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2099 **Figure captions**

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Fig. 1. Geological sketch map of the Dharwar Craton showing the three crustal blocks with strain fabrics and shear zone patterns (modified after [Chardon et al., 2008, 2011](#); [Peucat et al., 2013](#); [Jayananda et al., 2013a](#)).

Fig.2. Geological sketch map of the NW-SE crustal corridor (Hanagal – Madnapalle) displaying the different gneisses, greenstone sequences and granitoid plutons present in the area (modified after [Chardon et al., 2011](#); [Peucat et al. 2013](#)).

Fig.3. Representative field photographs of lithologies in the Central block: (a) Migmatitic gneiss in an abandoned quarry at the 46 km marker stone along the Challakere -Bellary road (near telecom tower). (b) Melt-filled shear zone traversing metapelite undergoing melting where large sillimanite crystals are rimmed by spinel-quartz. (c) Dark grey flat gneiss along the western boundary of the Closepet batholith near Rolla village. (d) Dark grey flat gneiss traversed by melt-filled shears, 2km east of Palasamudram along the Lepakshi -Kadiri road. (e) Quartz monzonite of the Sira magmatic complex with leucocratic veins extending along a flat magmatic foliation. (f) Coeval quartz monzonite and monzogranite showing diffuse contact with a biotite-rich rim and both cut by a later felsic vein, at 2 km west of Puttaparthi village. (g) Porphyritic monzogranite with MME at Reddicheruvukatta along the Gorantla-Kadiri road. (h) Anatectic granite dyke cutting across dark grey gneiss along the western boundary of the Closepet batholith (10 km south of Hangal). (i) Anatectic granite containing large angular remnants of mafic lower crust along the shear zone at the western margin of the Closepet batholith, about 1 km south of Kundurpi village. (j) Syn-plutonic mafic dyke defining magmatic flow in anatectic granite, about 8 km north of Madakasira along the Pavagada road.

Fig.4. Representative field photographs of lithologies in the Eastern block: (a) Banded dark grey transitional TTG gneiss at about 3 km west of Vepuraka railway station. (b) Diatexitic gneiss 3 km north of Horsely Hills Cross along the Madanapalle - Kadiri road. (c) Large quarry displaying interlayered dark grey banded transitional TTG, porphyritic monzogranite and diatexite, 5-km north of Madanapalle along the Kadiri road. (d) Dark grey monzodiorite with MME, 1 km north of Billur along the Bagepalli road. (e) Diatexitic gneisses together with quartz-monzonite define magmatic flow with biotite-rich rim along the contact (4 km marker stone along the Gollapalli-Muvedu road). (f) Fresh exposure displaying magma mingling with hybrid enclaves forming pillow-like structures (1 km south of Tanakallu along the Madanapalle-Kadiri road).

Fig. 5. Interpretative cross-section of the studied corridor from Hangal to Madanapalle showing lithological associations and fabrics.

Fig. 6. Photomicrographs of representative mineral assemblages from the Central block (crossed polars). (a) Quartz-plagioclase-K-feldspar-biotite assemblage in older migmatitic gneiss from the western boundary of the Closepet batholith (exposure near telecom tower). (b) Quartz-plagioclase-K-feldspar-hornblende-biotite in dark grey banded gneiss at Ravadi quarry (c) Quartz monzonite with quartz-plagioclase-K-feldspar-clinopyroxene-hornblende-biotite from Pavagada town quarry. (d) Monzogranite showing quartz-plagioclase-K-feldspar-hornblende-biotite from 2 km west of Puttaparthi village. (e) Syn-plutonic dyke from Madakasira showing quartz-plagioclase-K-feldspar-clinopyroxene-hornblende-biotite. (f)

2149 Anatectic granite showing bimodal quartz-K-feldspar-biotite at about 3 km SE of Pavagada
1 2150 along the Hindupur road. Mineral abbreviations: Kfl- K-feldspar, Plg-plagioclase, Cpx-
2 2151 clinopyroxene, Qtz-quartz; Bt, Bio-biotite. Amp -Amphibole.
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4 2153 Fig. 7. Photomicrographs of representative mineral assemblages from the Eastern block: (a)
5 2154 Dark grey banded gneiss from 2 km west of Vepuraka, showing quartz-plagioclase-K-
6 2155 feldspar-hornblende-biotite. (b) Quartz monzonite from the eastern margin of the Kadiri
7 2156 greenstone belt showing quartz-plagioclase-K-feldspar-hornblende-biotite. (c) Syn-plutonic
8 2157 dyke (near Tanakallu) showing quartz-plagioclase-K-feldspar-hornblende-biotite. (d) Quartz-
9 2158 K-feldspar-biotite in anatectic granite from Horsley Hills.
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11 2160 Fig. 8.1a-8.1d – U-Pb concordia diagrams of migmatitic TTG and dark grey banded gneisses
12 2161 from the Central block. CL images of representative analysed grains are also provided in the
13 2162 figures. 8.1a - U-Pb concordia diagram of zircons analysed from migmatitic TTG (EDC TTG
14 2163 2b); 8.1b - U-Pb concordia diagram of transitional TTGs from sample 17 CL01); 8.1c –
15 2164 Zircon U-Pb concordia diagram of dark grey banded gneiss (KO 32); 8.1d - Zircon U-Pb
16 2165 concordia diagram for transitional TTG (KO20). See the text for details.
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18 2167 Fig. 8.2a to 2f. U-Pb concordia diagrams for zircons analysed from sanukitoids, syn-plutonic
19 2168 mafic dyke and anatectic granite from the Central block. CL images of representative
20 2169 analysed zircon grains are also presented in the figures. 8.2a – U-Pb concordia diagram of
21 2170 zircons analysed from sanukitoid (PAV-VES 01a); 8.2b – U-Pb concordia diagram of zircons
22 2171 analysed from sanukitoid (KO18); 8.2c – U-Pb concordia diagram of zircons analysed from
23 2172 sanukitoid (KO19a); 8.2d – U-Pb concordia diagram of zircon analysed from granitoid
24 2173 (KO15); 8.2e – U-Pb concordia diagram of zircon analysed from synplutonic dyke (PAV-
25 2174 VES 01d); 8.2f – U-Pb concordia diagram of zircon analysed from anatectic granite (KO22).
26 2175 See the text for details.
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28 2177 Fig.8.3a to 3d. U-Pb concordia diagrams for transitional TTGs, sanukitoids and anatectic
29 2178 granite from the eastern block. CL images of representative zircon grains are also provided in
30 2179 the figures. 8.3a – U-Pb Concordia diagram of zircon analysed from transitional TTGs
31 2180 (KO8); 8.3b – U-Pb concordia diagram of zircon analysed from transitional TTGs (KO7);
32 2181 8.3c – U-Pb concordia diagram of granitoid from sample (KO10); 8.3d – U-Pb concordia
33 2182 diagram of granitoid from sample (KO29a). See the text for details.
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35 2184 Fig. 9 (a). Normative Ab-An-Or triangular diagram (after [O'Connor, 1965](#); updated by
36 2185 [Barker, 1979](#)) with fields from [Barker \(1979\)](#); (b). Na-K-Ca triangular plot (after [Barker and](#)
37 2186 [Arth, 1976](#)) discriminating the calc-alkaline differentiation trend of sanukitoids, anatectic
38 2187 granite from the trondhjemite trend of the TTGs. It also shows the transitional character of
39 2188 the younger transitional TTGs; (c). $Al_2O_3 / (CaO + Na_2O + K_2O \text{ mol})$ versus K_2O / Na_2O plot
40 2189 showing the samples plotting in the fields of TTG, sanukitoid and biotite granite; (d).
41 2190 $Al_2O_3 / (FeO^I + MgO)$ versus $CaO / (Na_2O + K_2O)$ plot (after [Laurent et al., 2014](#)) showing the
42 2191 transitional character of transitional TTGs with respect to the sanukitoids and biotite-granite.
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44 2193 Fig. 10. Key major element Harker diagrams for transitional TTGs, sanukitoids, anatectic
45 2194 granites, and syn-plutonic dykes.
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47 2196 Fig.11. Chondrite normalized and primitive mantle normalized diagrams for rocks from the
48 2197 Central block. (a) Chondrite normalized REE patterns of transitional TTGs; (b) Primitive
49 2198 mantle normalized multi-element spider diagram ([Sun and Mc Donough, 1989](#)) of transitional
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2199 TTGs; (c) Chondrite normalized REE patterns of sanukitoids; (d) Primitive mantle
1 2200 normalized multi-element spider diagram of sanukitoids; (e) Chondrite normalized REE
2 2201 patterns of anatectic granites; (f) Primitive mantle normalized spider diagram of anatectic
3 2202 granites; (g) Chondrite normalized REE patterns of group 1 syn-plutonic mafic dykes; (h)
4 2203 Primitive mantle normalized multi-element spider diagram of group 1 syn-plutonic dykes; (i)
5 2204 Chondrite normalized REE patterns of group 2 syn-plutonic dykes; (j) Primitive mantle
6 2205 normalized multi-element spider diagram of group 2 syn-plutonic dykes.
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9 2207 Fig.12. Chondrite normalized and primitive mantle normalized diagrams for rocks from the
10 2208 Eastern block. (a) Chondrite normalized REE patterns of transitional TTGs; (b) Primitive
11 2209 mantle normalized multi-element spider diagram of transitional TTGs; (c) Chondrite
12 2210 normalized REE patterns of sanukitoids; (d) Primitive mantle normalized multi-element
13 2211 spider diagram of sanukitoids; (e) Chondrite normalized REE patterns of anatectic granites;
14 2212 (f) Primitive mantle normalized spider diagram of anatectic granites; (g) Chondrite
15 2213 normalized REE patterns of synplutonic mafic dykes; (h) Primitive mantle normalized multi-
16 2214 element spider diagram of syn-plutonic dykes.
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19 2216 Fig.13. (a) Initial $^{176}\text{Hf}/^{177}\text{Hf}$ versus zircon ages plot for the transitional TTGs, sanukitoids
20 2217 syn-plutonic mafic dykes and anatectic granites; (b) $\epsilon\text{Hf}_{(T)}$ versus zircon ages obtained in this
21 2218 study and from published data (Bidyananda et al., 2016); (c) $\epsilon\text{Nd}_{(T)}$ versus zircon ages of
22 2219 rocks from the present study and from published data (Krogstad et al., 1995; Peucat et al.,
23 2220 1993; Jayananda et al., 2000; Moyen, 2000; Dey et al., 2014).
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26 2222 Fig. 14. (a) $3*\text{CaO}-\text{Al}_2\text{O}_3/(\text{FeO}^{\text{t}}+\text{MgO})-5*(\text{K}_2\text{O}/\text{Na}_2\text{O})$ triangular plot (Laurent et al., 2014)
27 2223 depicting the sources of transitional TTGs, sanukitoids, syn-plutonic dykes and anatectic
28 2224 granites; (b) Sr+Ba versus $\text{FeO}^{\text{t}}+\text{MgO}$ plot (Laurent et al., 2014) showing enrichment of
29 2225 sources from transitional TTGs to sanukitoids.
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32 2227 Fig.15. (a) Sr/Y versus Y binary plot (after Defant and Drummond, 1990) showing the role of
33 2228 mineral phases in the source residue. (b) $(\text{La}/\text{Yb})_{\text{N}}$ versus $(\text{Yb})_{\text{N}}$ plot (Martin, 1986)
34 2229 characterizing residual garnet or amphibole. The majority of samples plot in the area without
35 2230 residual garnet or with only 7-15 % residual garnet.
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38 2232 Fig. 16. Nb/Ta versus Zr/Sm plot (Foley et al., 2002) showing residual amphibole and
39 2233 ilmenite in an arc source residue, but without any rutile.
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42 2235 Fig. 17. Summary of zircon age data from this study and published data to highlight Archean
43 2236 crustal evolution in the Central and Eastern Dharwar Craton.
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46 2238 Fig.18. $(\text{Hf}/\text{Sm})_{\text{N}}$ versus $(\text{Ta}/\text{La})_{\text{N}}$ plot (La Flèche et al., 1998) showing dominantly melt-
47 2239 related enrichment, with minor fluid-related enrichment, of the mantle sources involved in
48 2240 generation of the transitional TTGs, sanukitoids and syn-plutonic dykes.
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51 2242 Fig.19. Model showing westward convergence of hot oceanic lithosphere during the
52 2243 assembly of the micro-blocks. Eventual slab break-off led to asthenosphere upwelling
53 2244 inducing widespread mantle melting resulting in juvenile magmatic additions resulting from
54 2245 lateral flow of ductile hot orogenic crust on hot weaker mantle. This led to 3D mass
55 2246 distribution and formation of orogenic plateaux with subdued topography as indicated by
56 2247 strain fabric data and strong seismic reflectivity along the E-W crustal profile.
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2249 **Supplementary Data**

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Appendix: Analytical methods for zircon U-Pb geochronology

Supplementary table 1. Zircon U-Pb data for the transitional TTGs, sanukitoids, syn-plutonic dykes and anatectic granites.

Supplementary table 2. *In situ* Hf isotope data for the transitional TTGs, sanukitoids, syn-plutonic dykes and anatectic granites.

Supplementary table 3. Major and trace element data for transitional TTGs, sanukitoids, syn-plutonic dykes and anatectic granites.

Supplementary table 4. Results for whole-rock Sm–Nd isotope measurements by thermal ionization mass spectrometry (TIMS).

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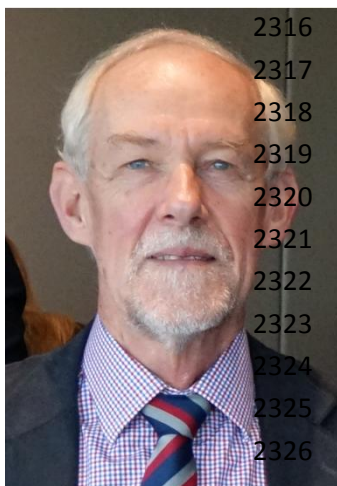
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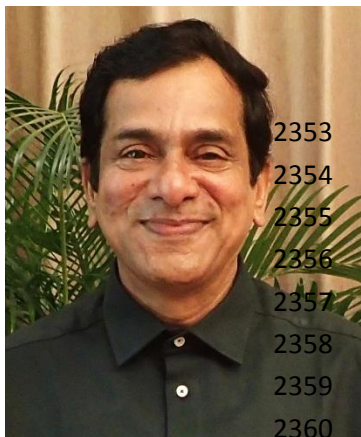
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General of the International Association for Gondwana Research. Research fields include petrology, fluid inclusions, geochemistry, geochronology, metallogeny and supercontinent tectonics. Published over 1200 research papers, edited several memoir volumes and journal special issues, and co-authored the book 'Continents and Supercontinents' (Oxford University Press, 2004). Recipient of National Mineral Award, Outstanding Geologist Award, Thomson Reuters 2012 Research Front Award, and Thomson Reuters/Clarivate Analytics High Cited Researcher 2014, 2015, 2016, 2017.



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R.V. Gireesh is a geologist in Karnataka state department of Mines and Geology. He has served as an Assistant Professor in Suratkal University and Central University of Karnataka. He has obtained MSc degree from University of Mysore and PhD from University of Bangalore. He has worked Neoproterozoic calc-alkaline magmatism with special focus on magma chamber dynamics and interaction of coeval felsic and mafic magmas. His interest include magmatism with associated mineralization in different tectonic settings.

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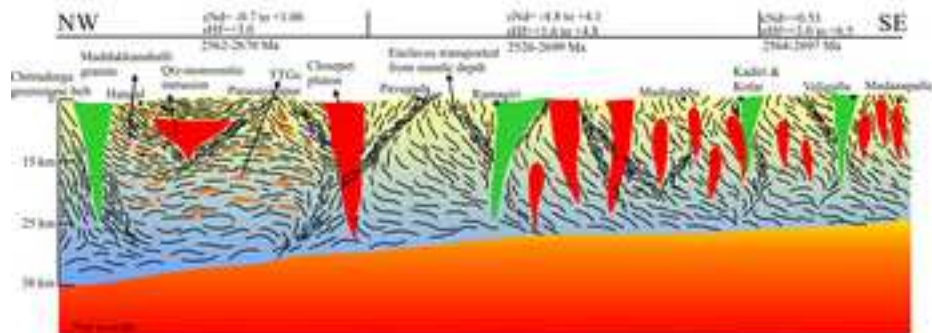


Figure: Interpretative cross-section of the studied corridor.

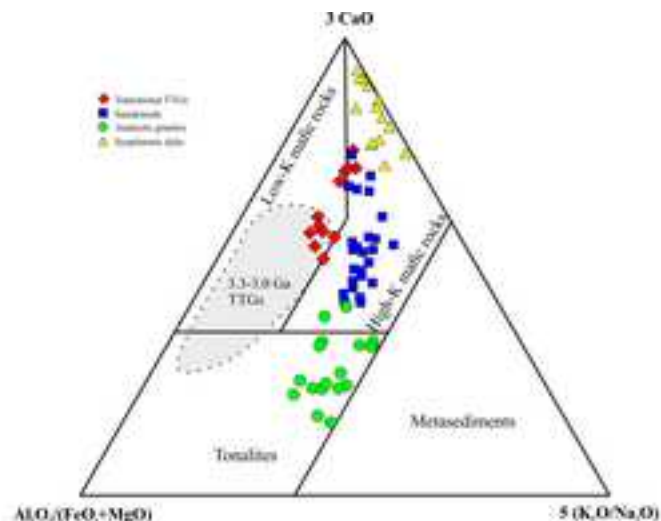


Figure: Ternary diagram (Laurent et al 2014) indicating the source rocks of gneisses and granitoids.

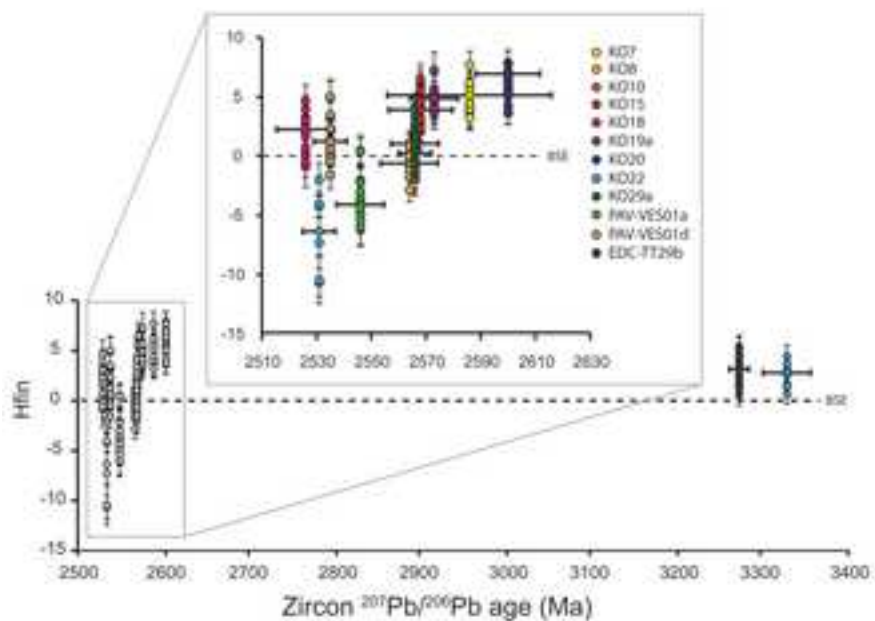


Figure: $\epsilon\text{Hf}(t)$ vs zircon ages indicating source mantle and crustal signatures.

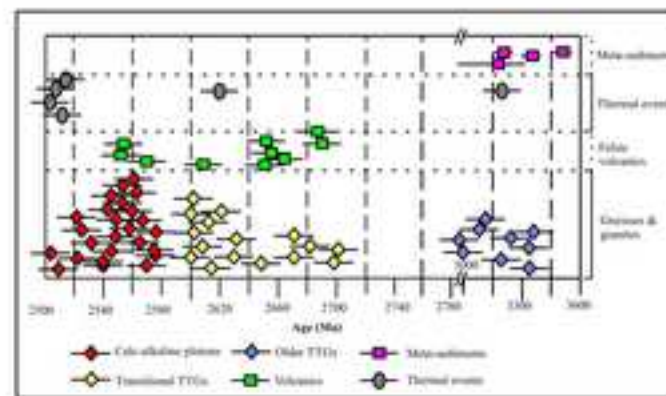
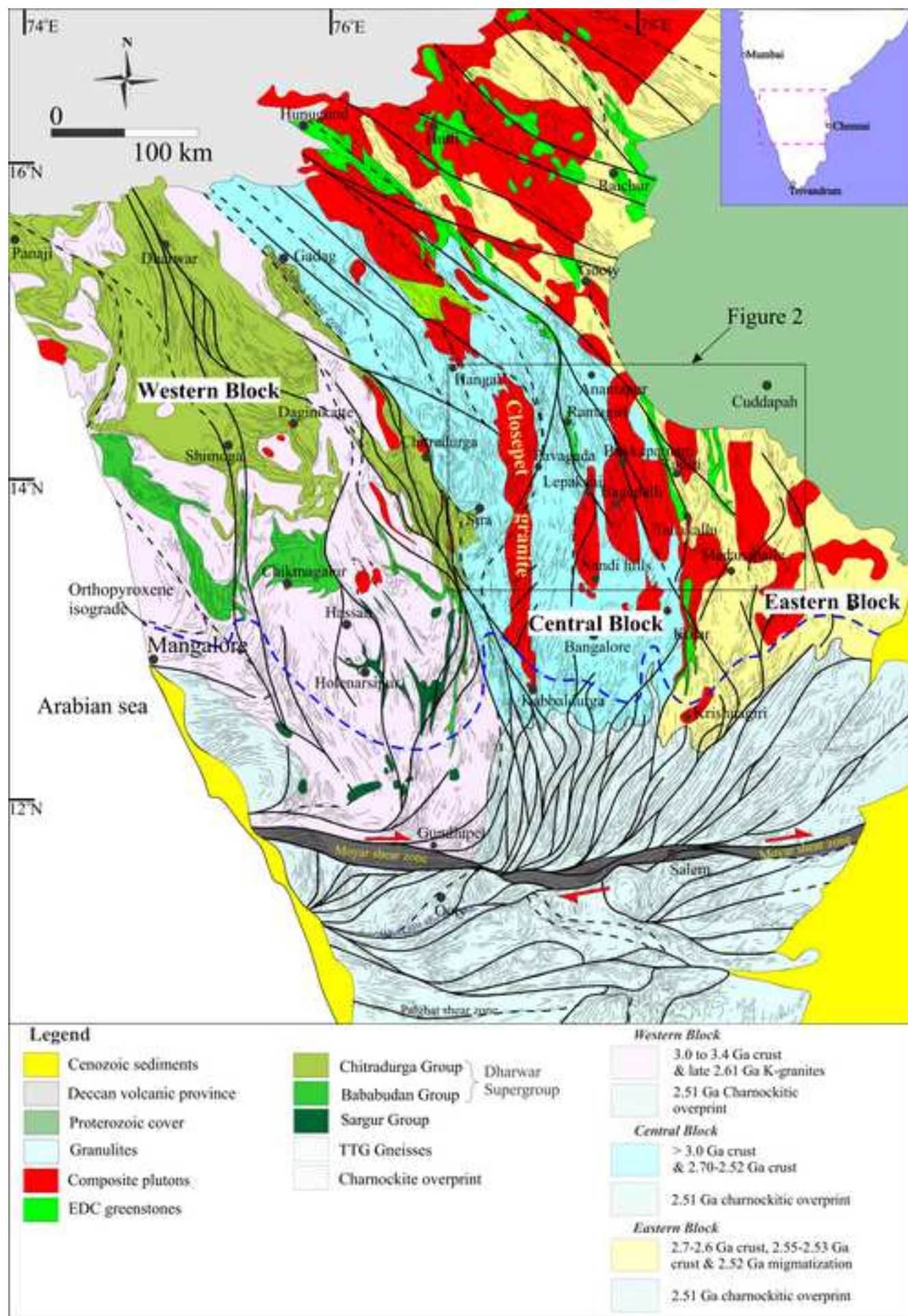


Figure showing multi-stage crustal growth, thermal events and cratonization.

Research Highlights

- Multi-stage Archean crustal growth ca. 3360-3200 Ma, ca. 3000-2960 Ma, ca. 2700-2600 Ma and ca. 2570-2520 Ma.
- Secular changes in the composition of granitoids through Neoproterozoic suggest increasing involvement of enriched peridotitic mantle.
- Convergence of oceanic lithosphere and eventual slab breakoff lead to asthenosphere upwelling.
- Lateral flow of hot orogenic crust lead to orogenic plateau formation with subdued topography.

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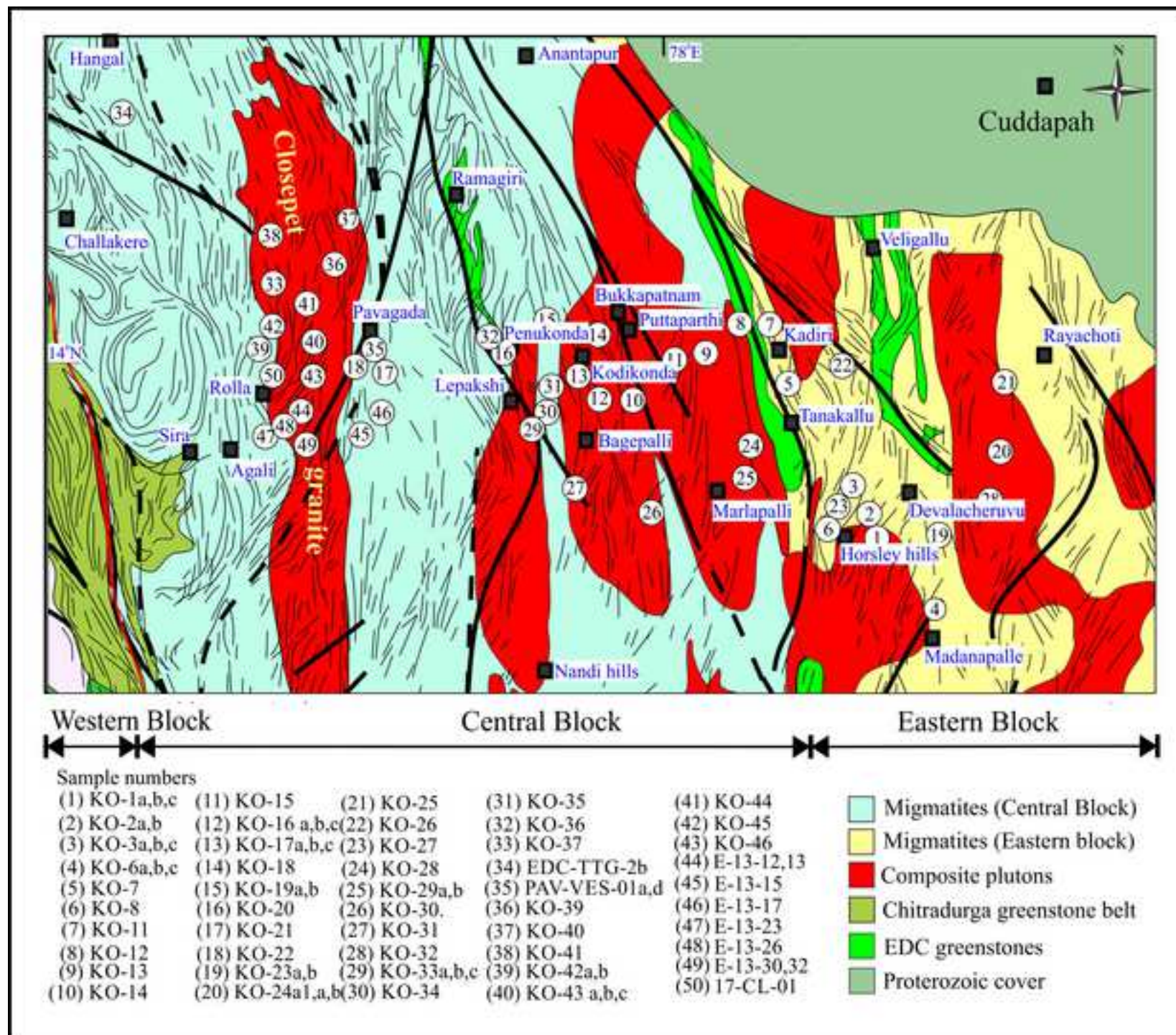


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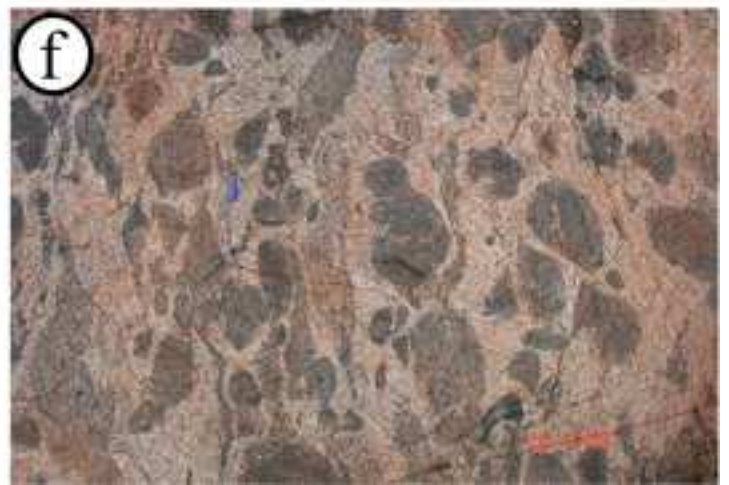


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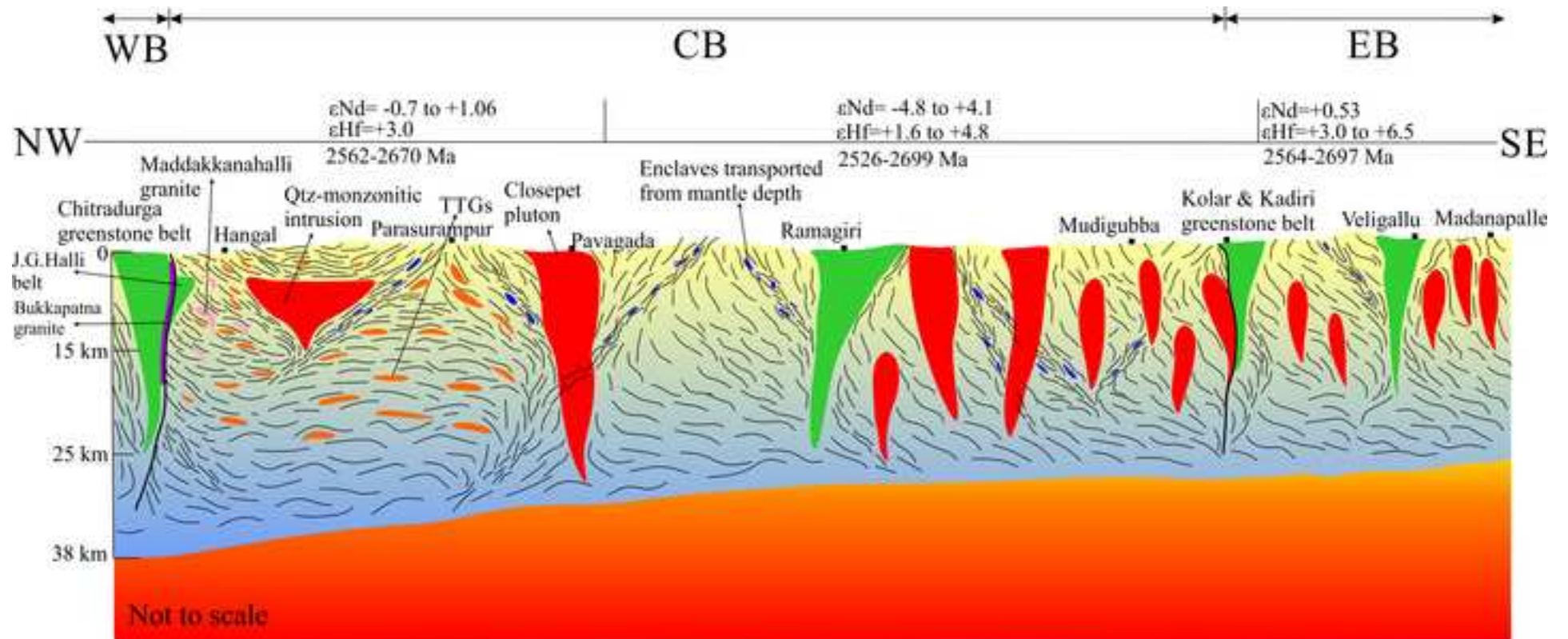


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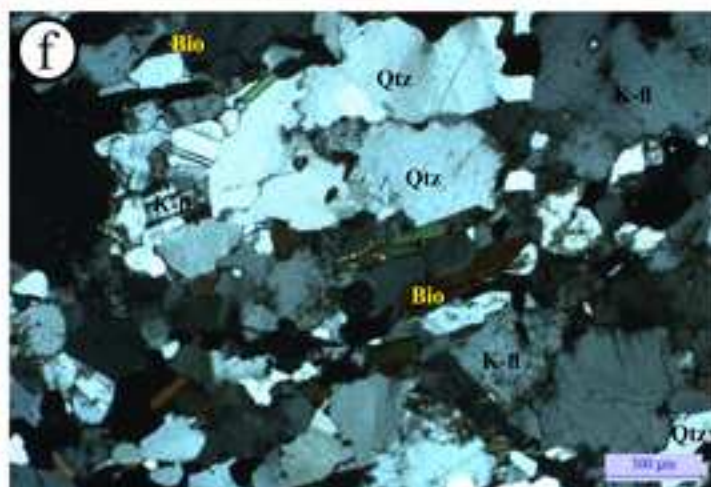
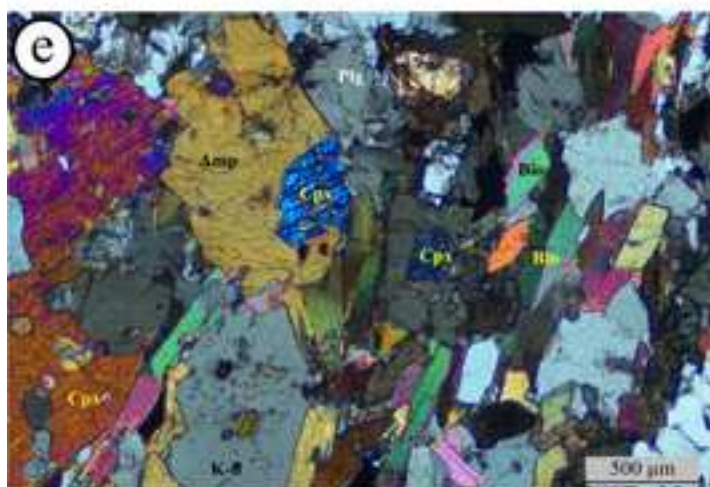
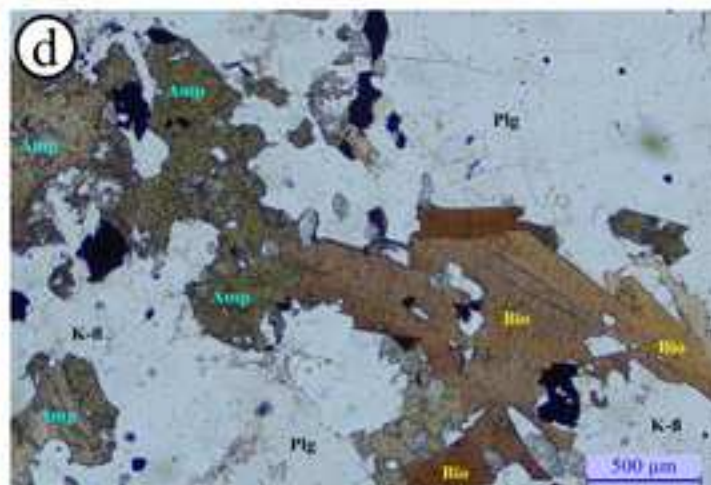
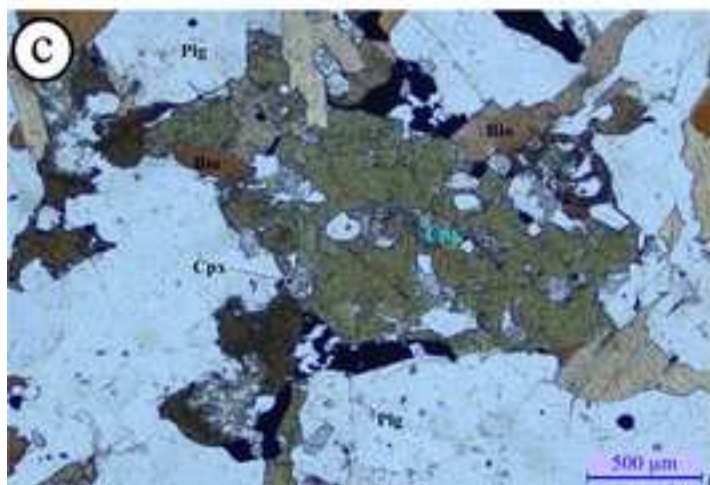
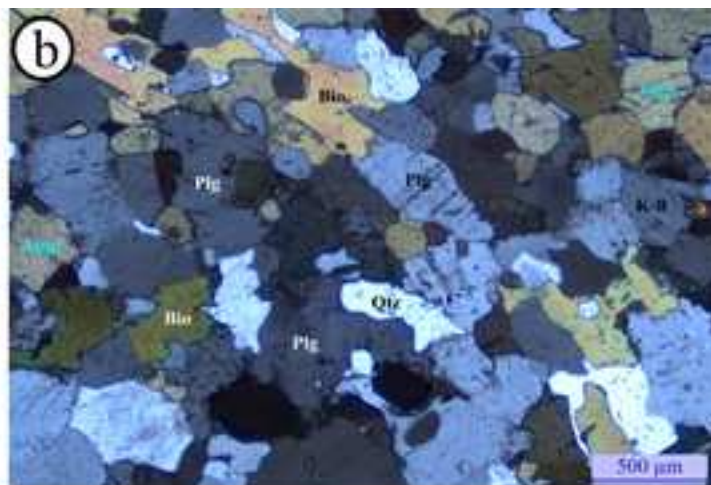
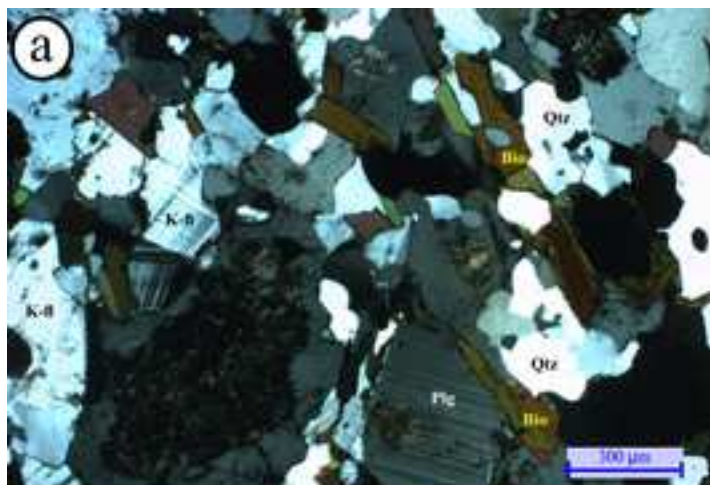
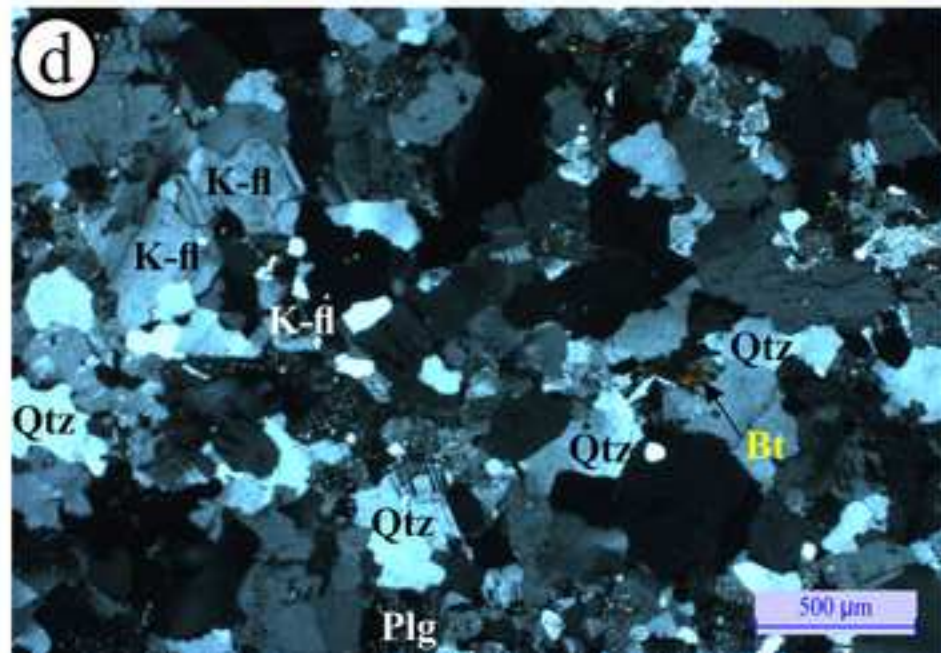
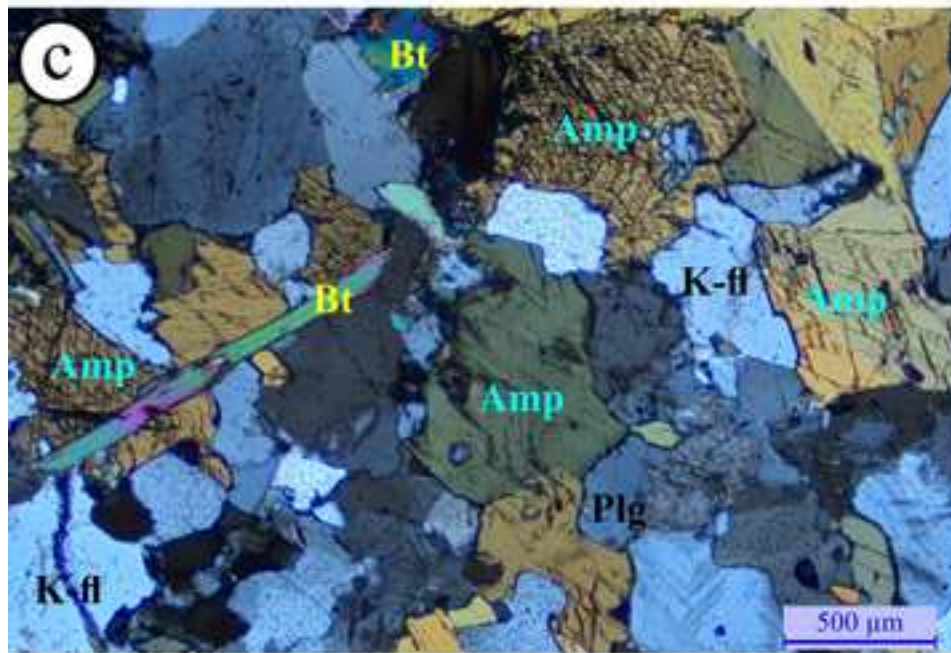
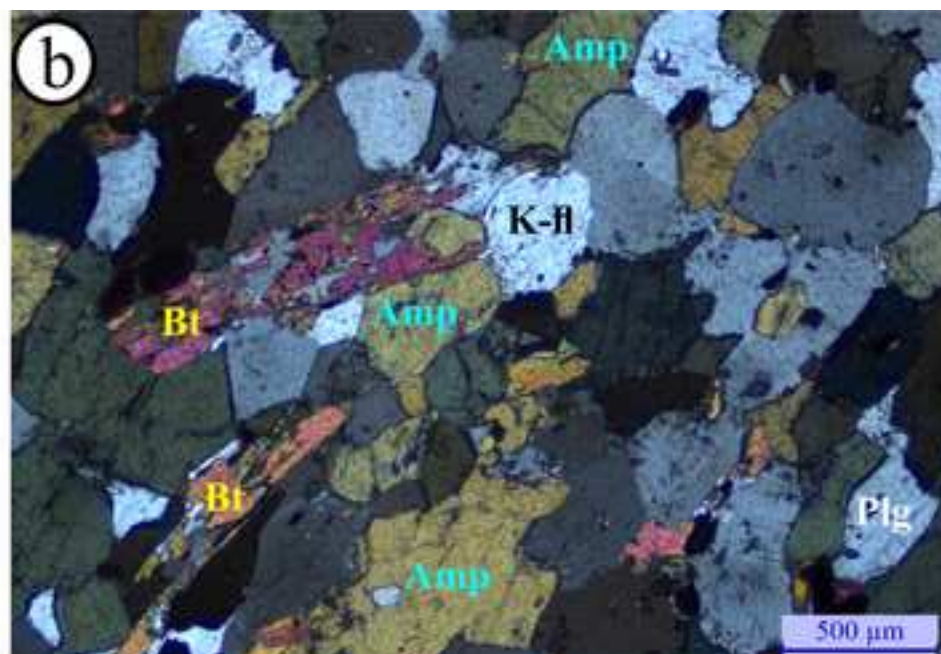
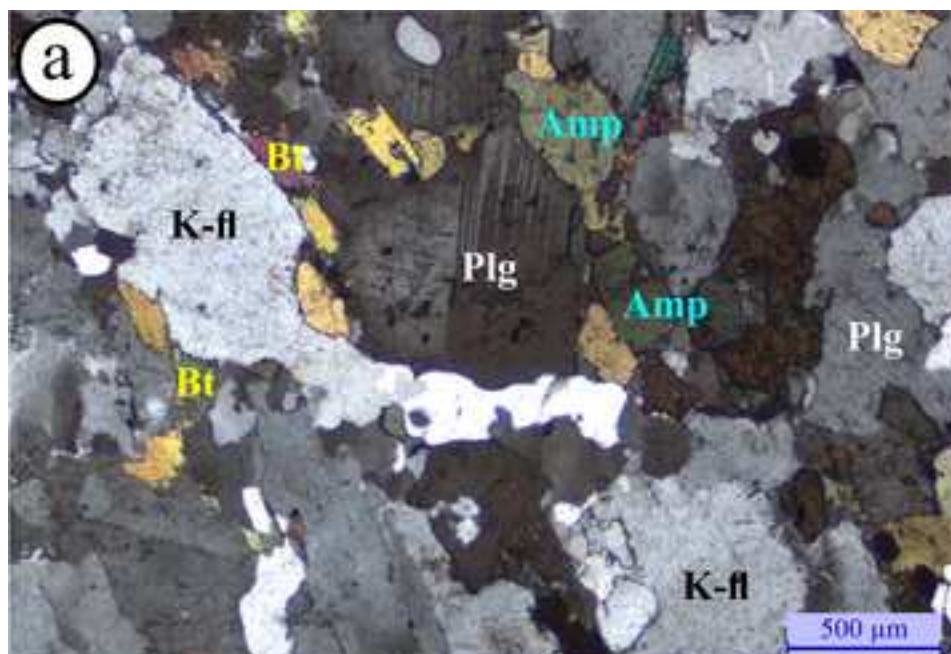
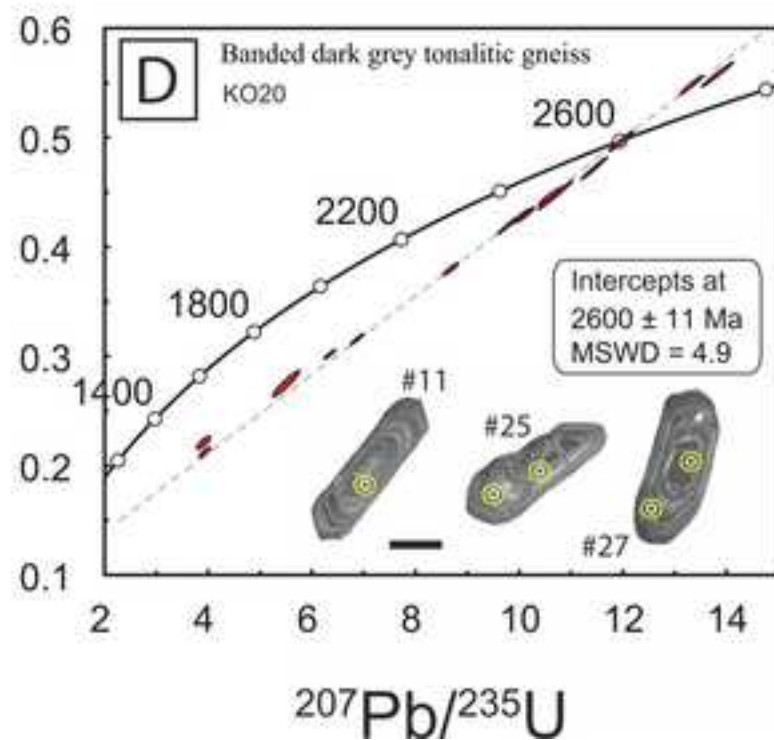
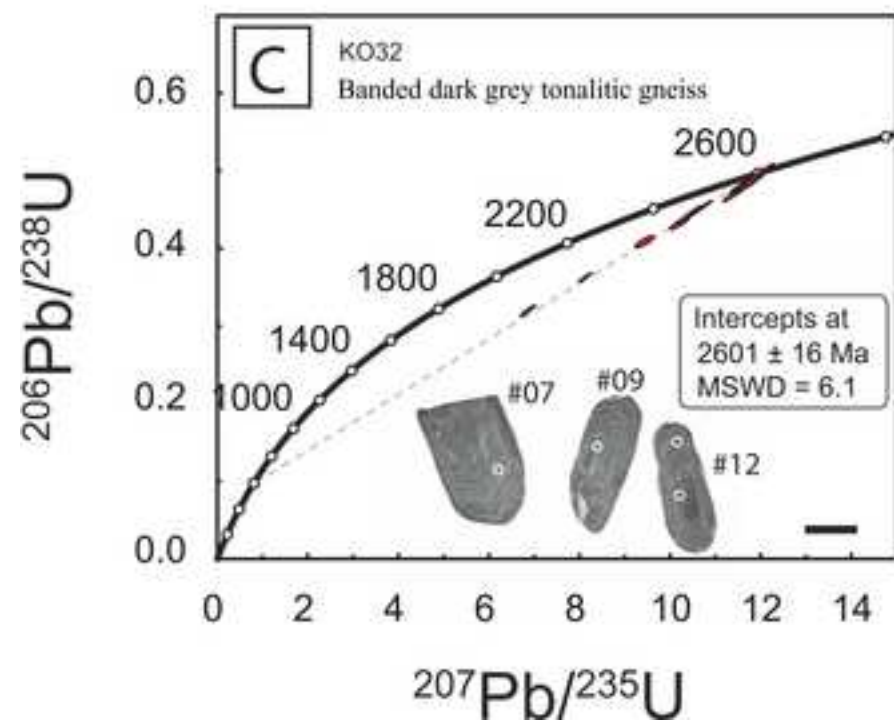
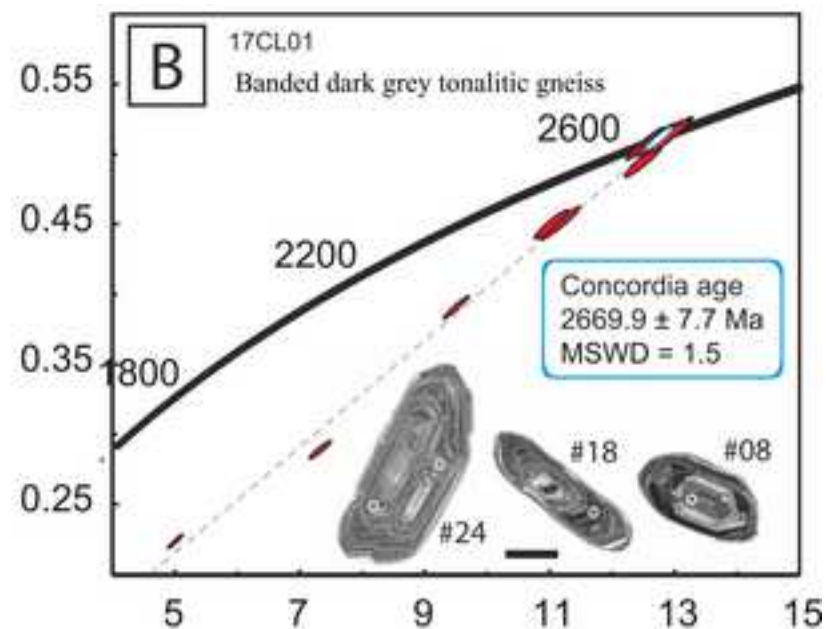
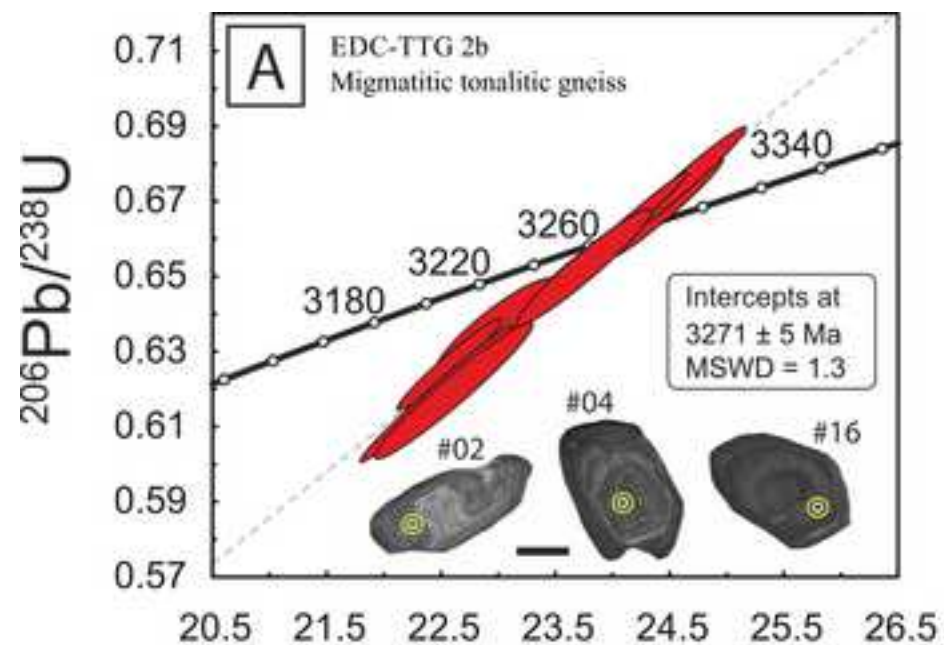


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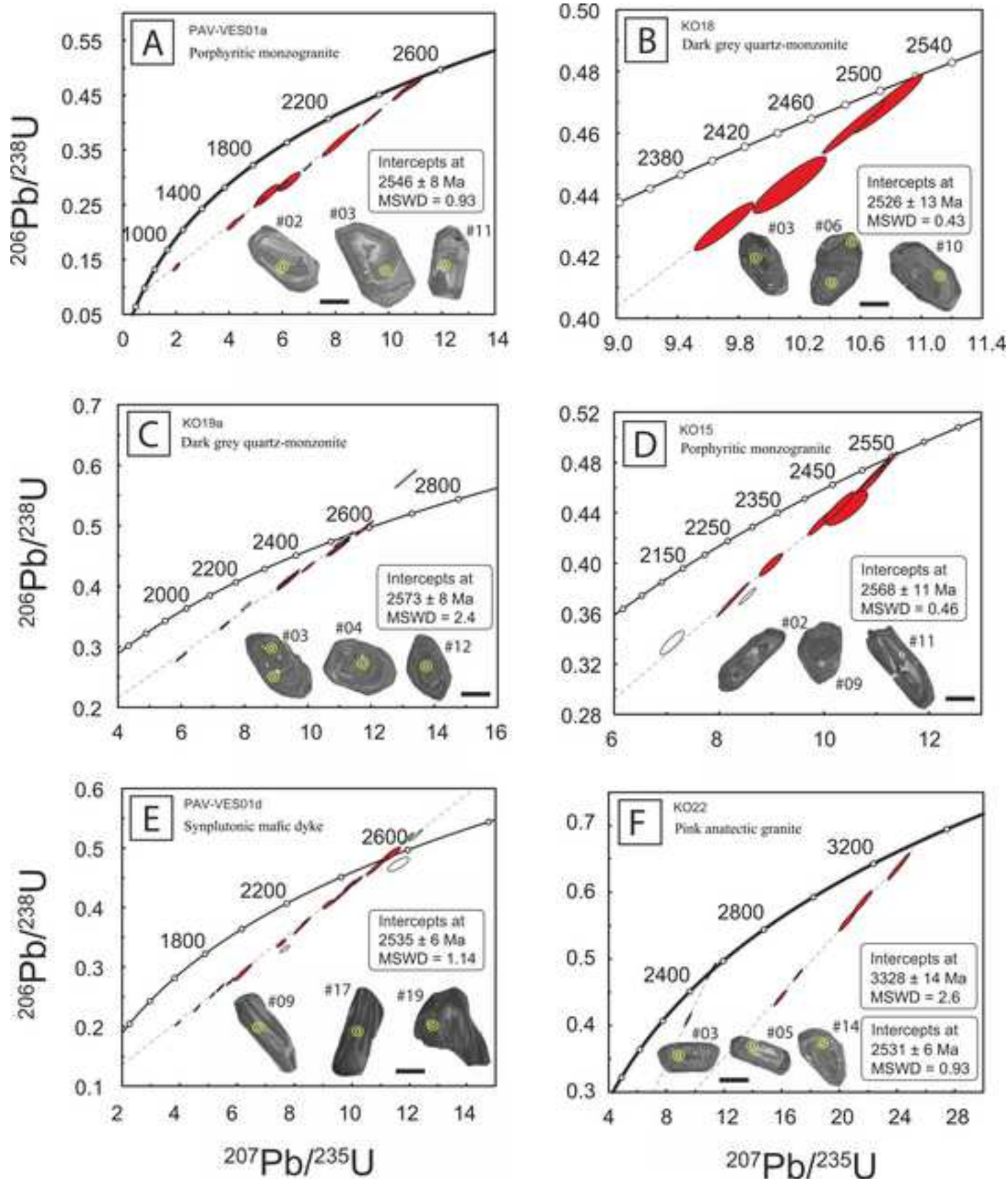
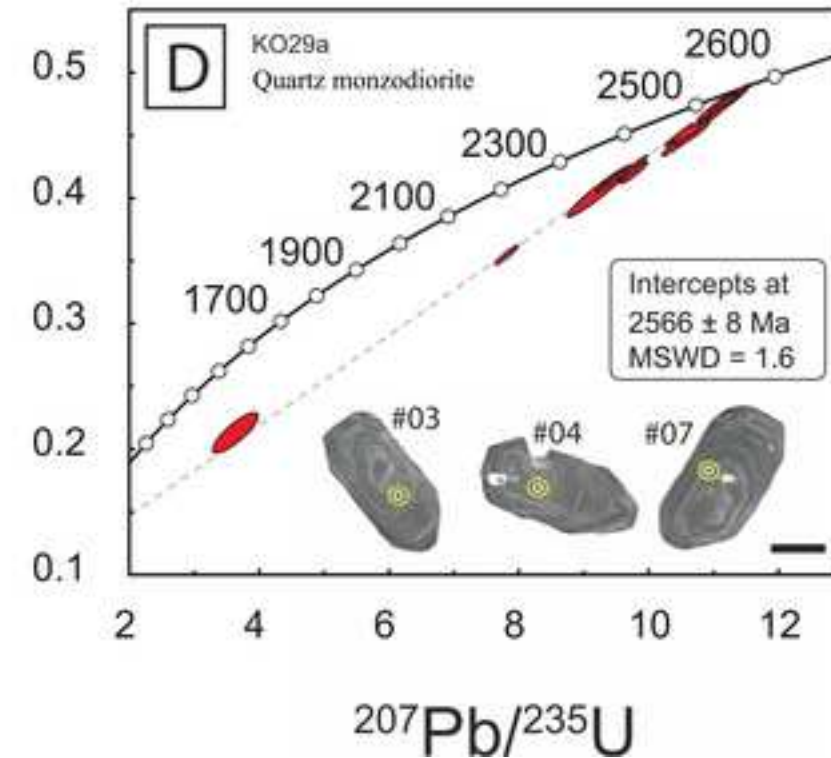
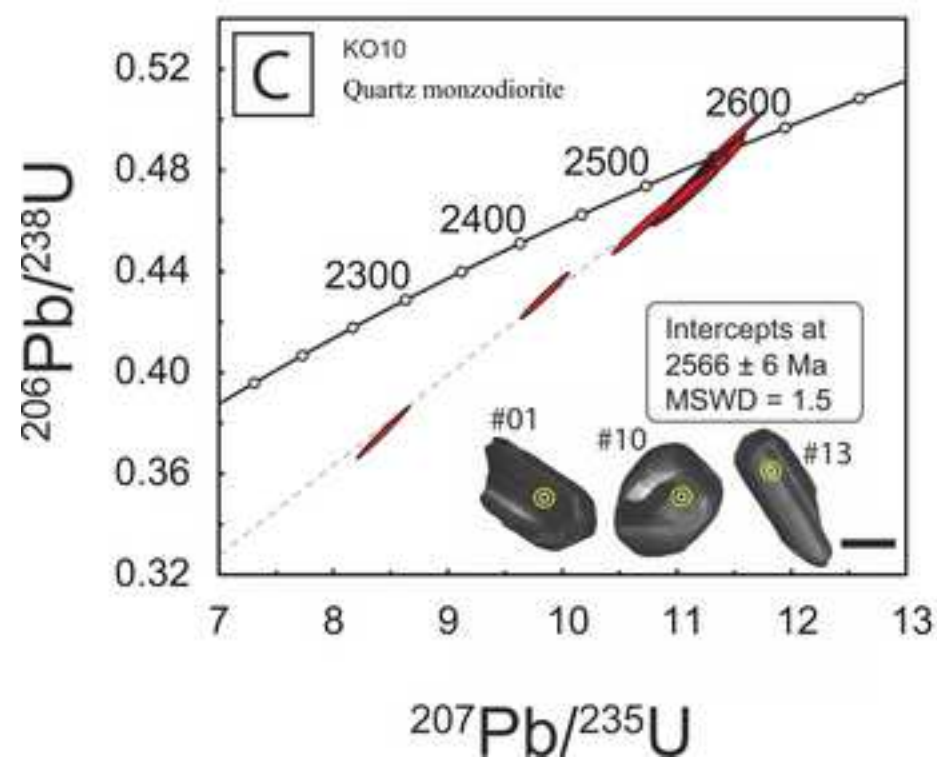
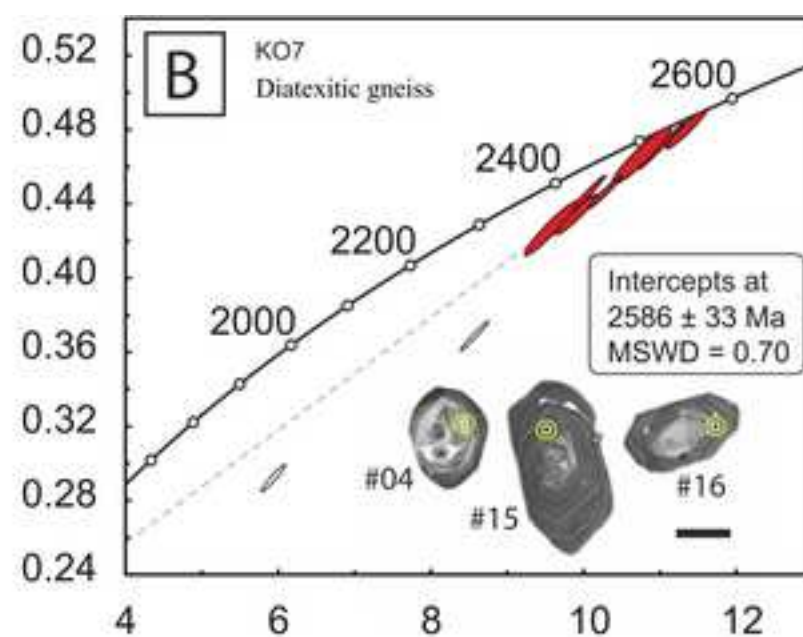
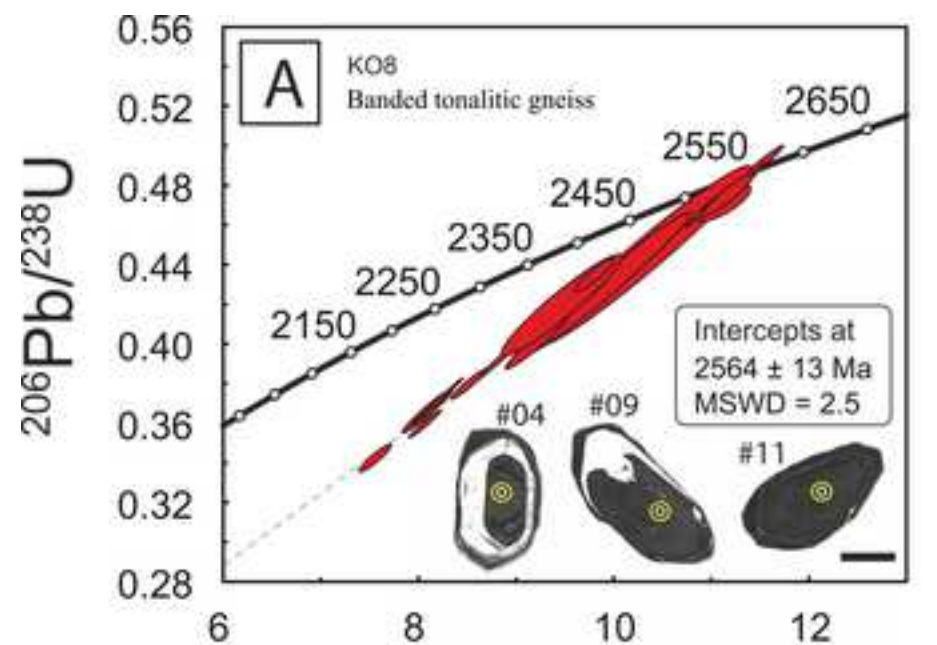


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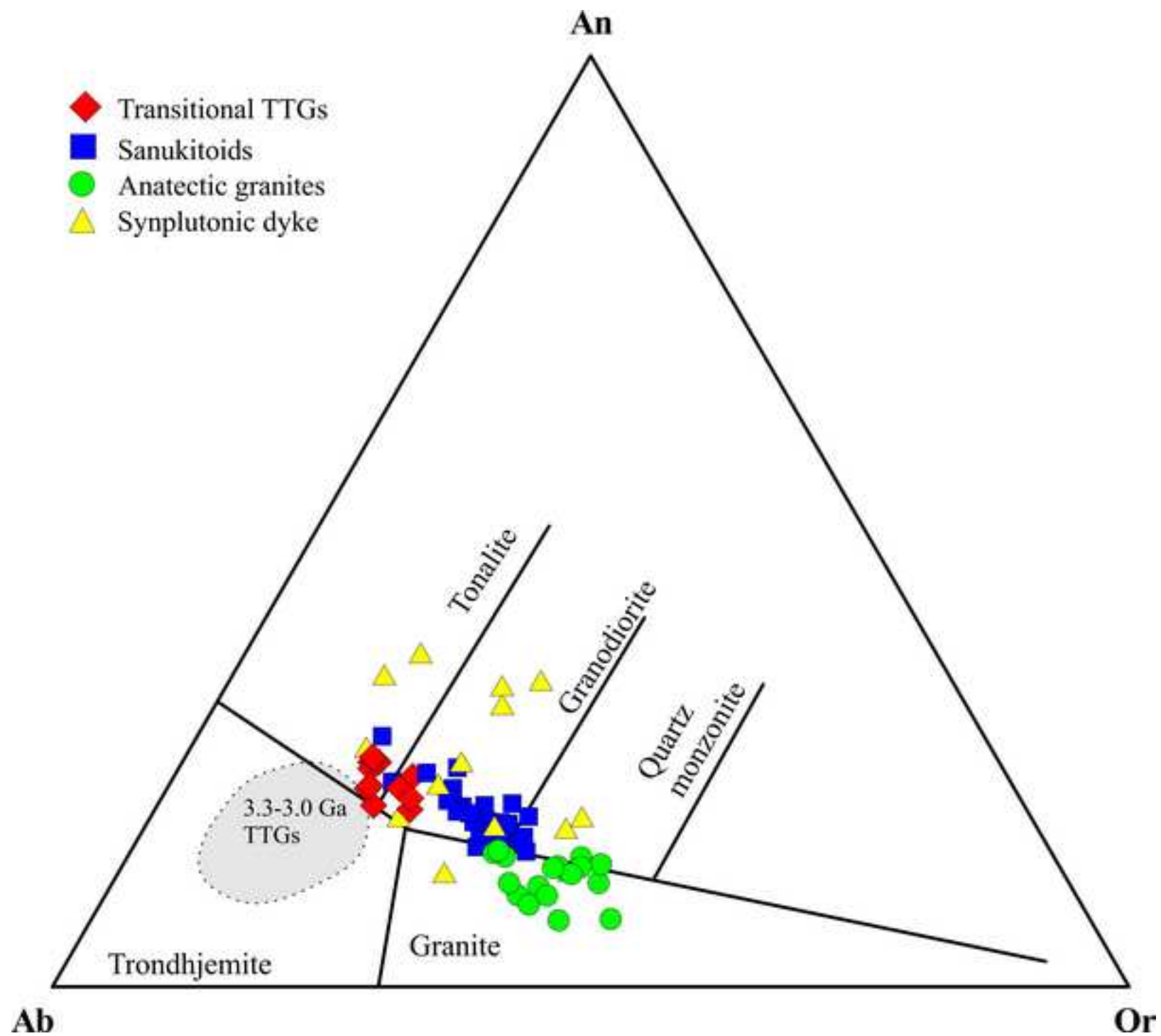


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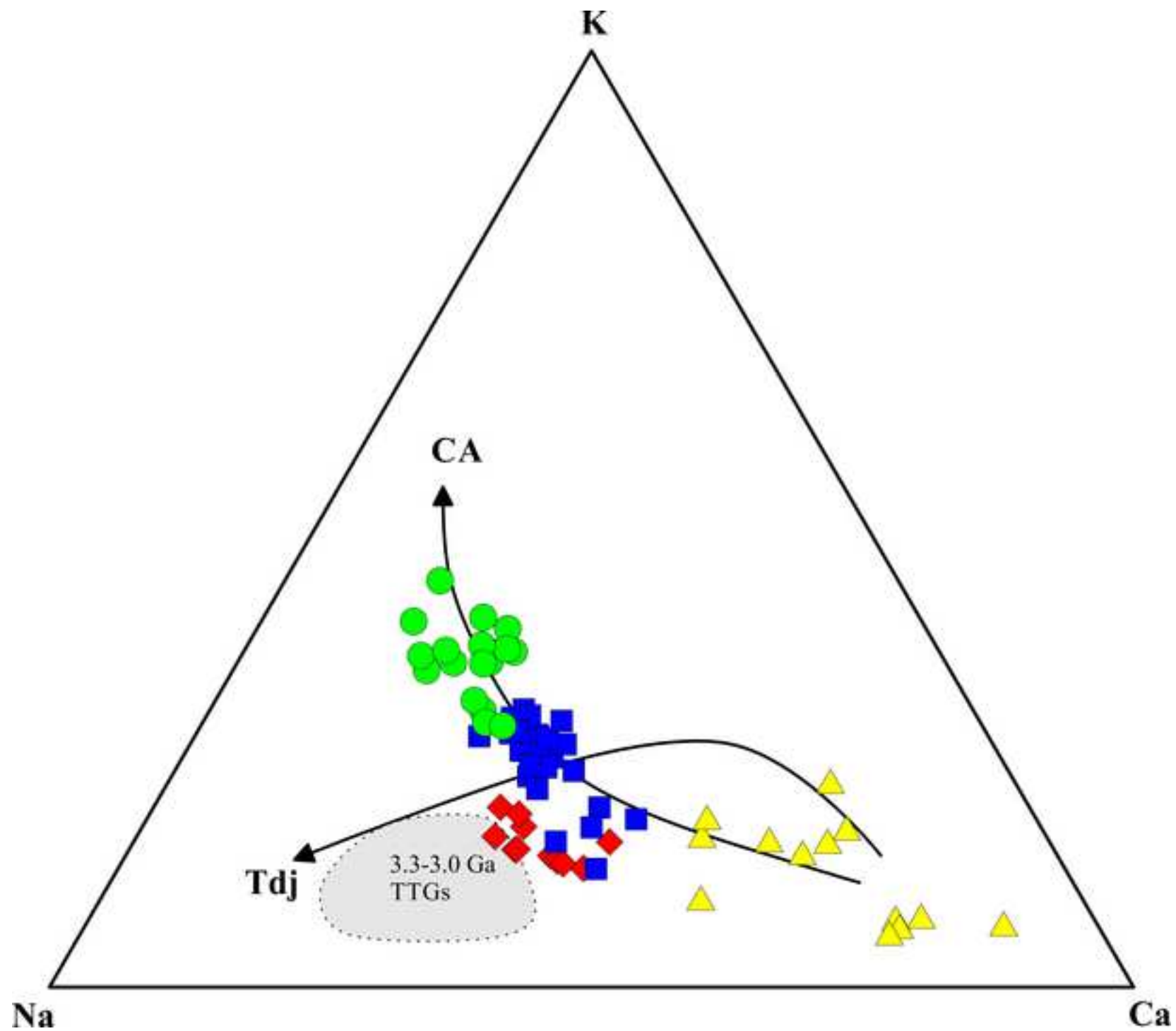


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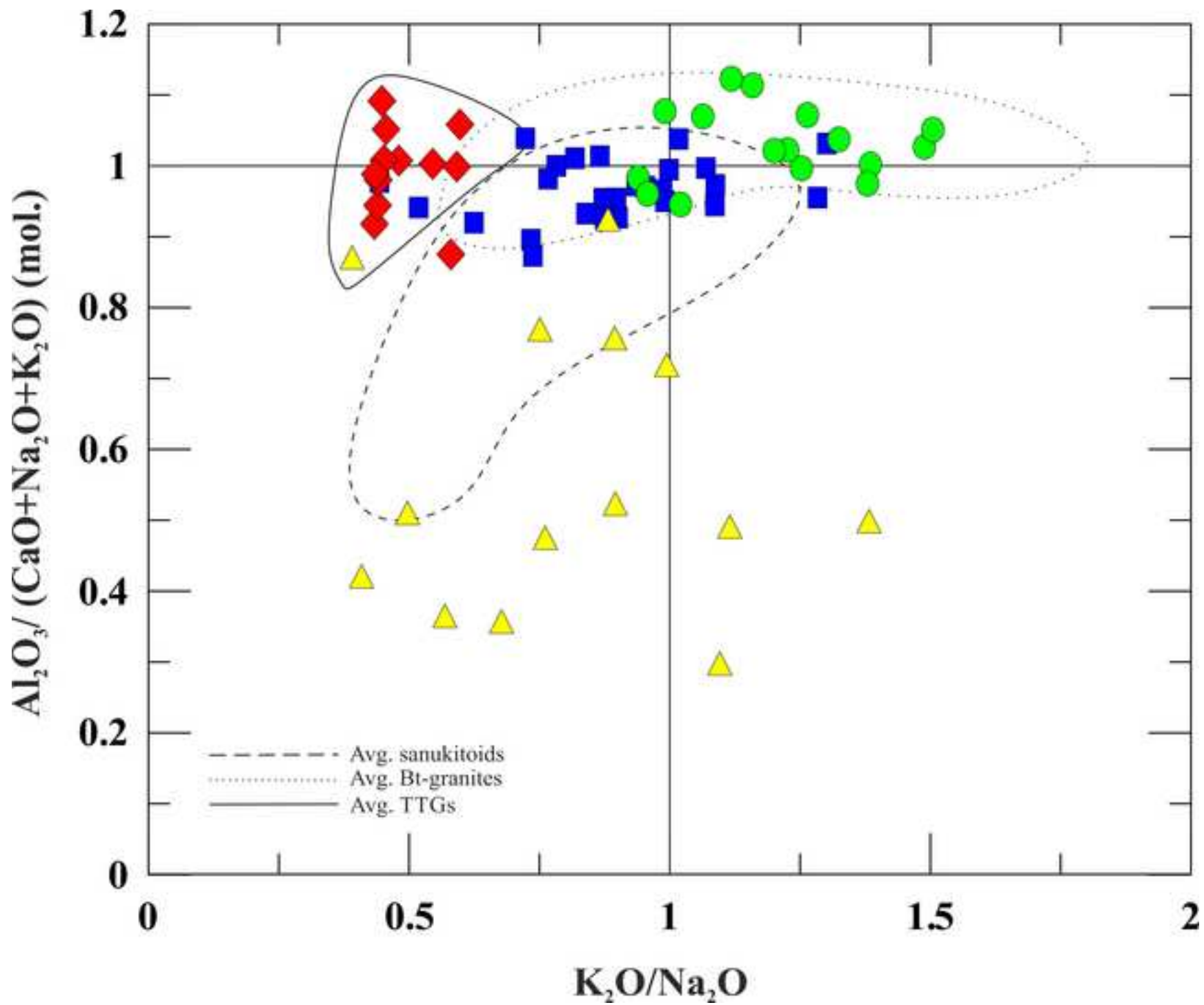


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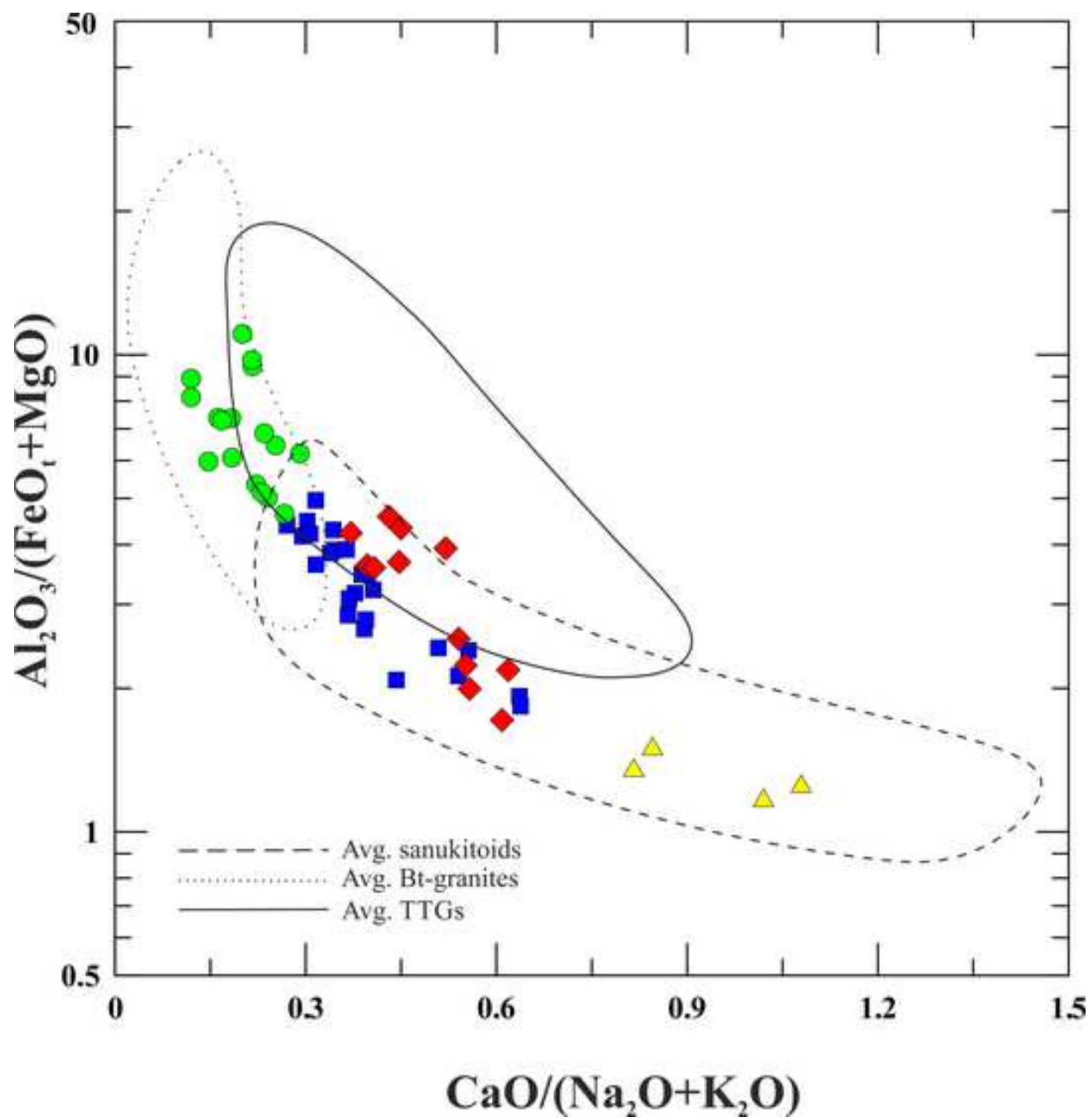


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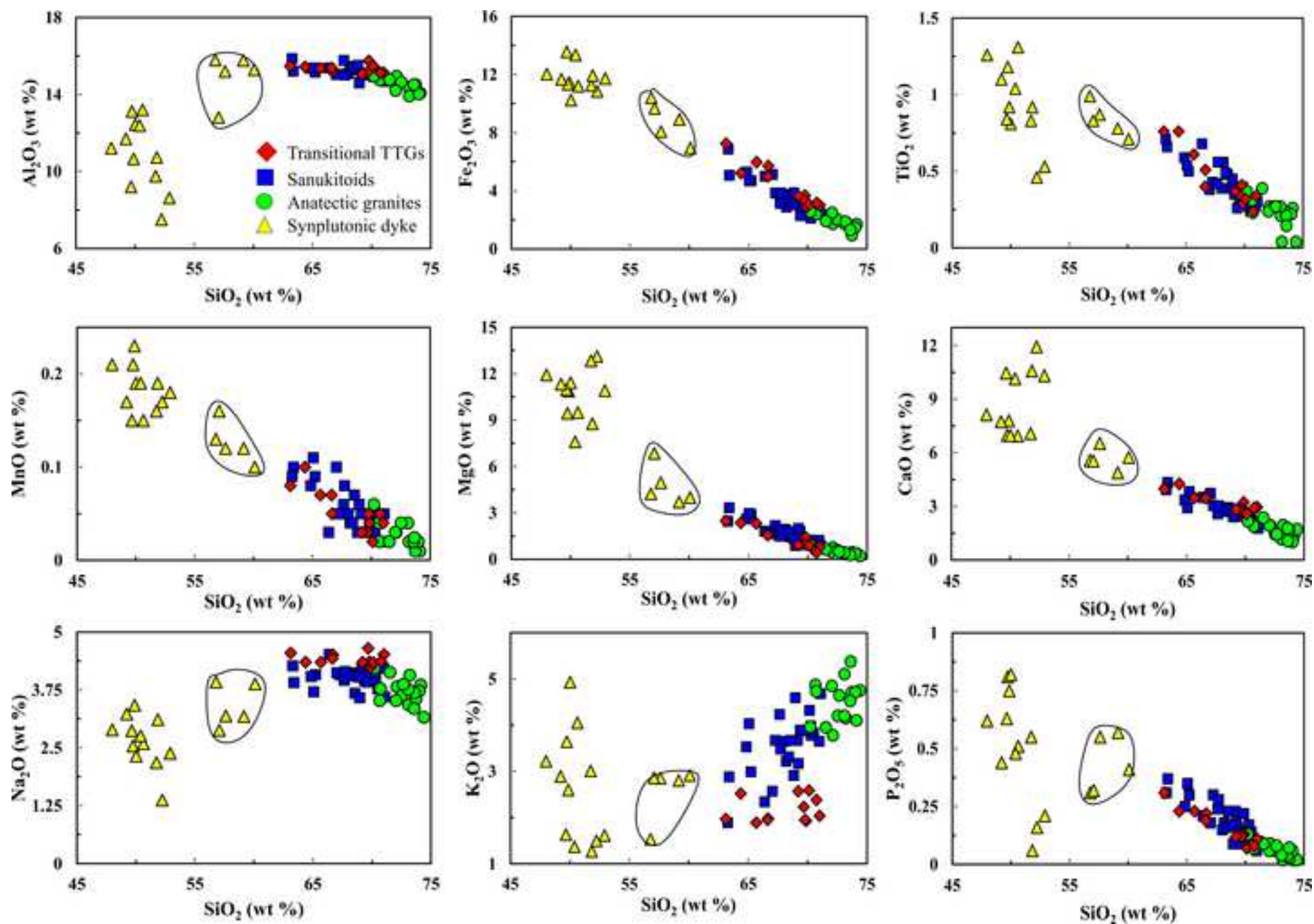


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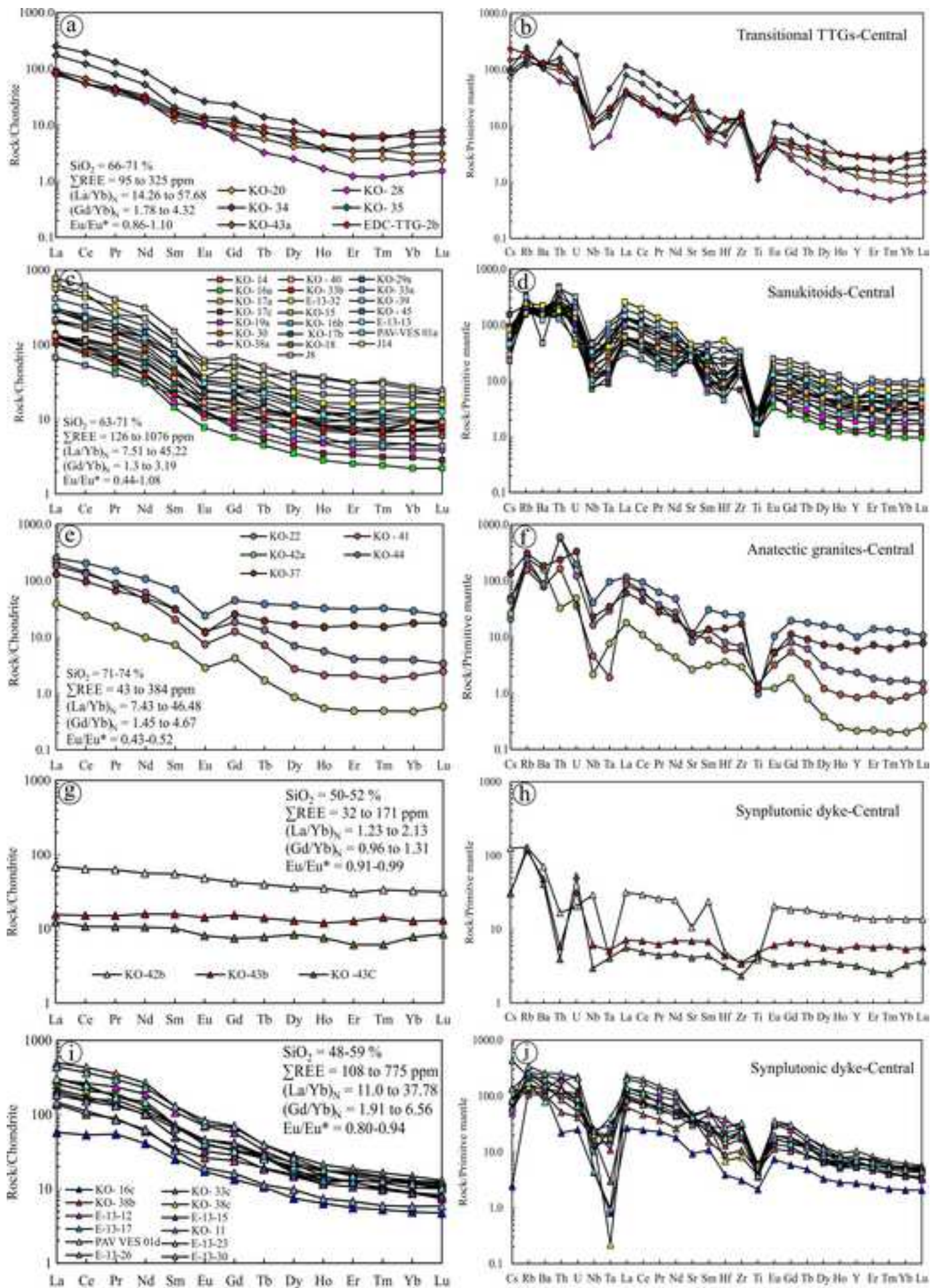


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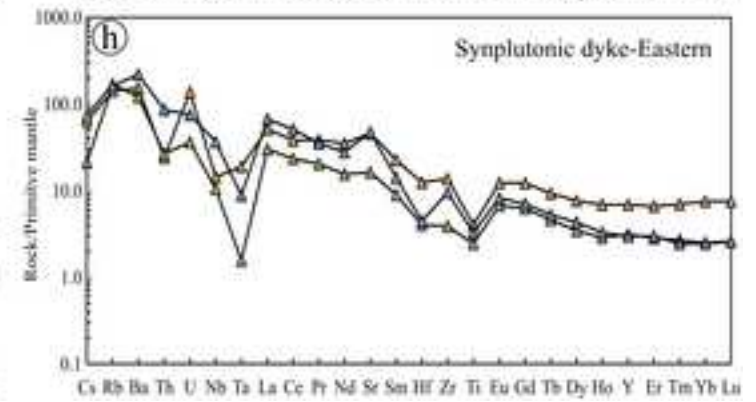
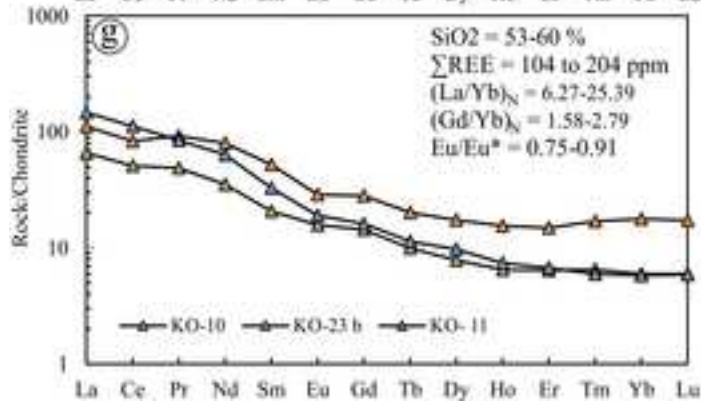
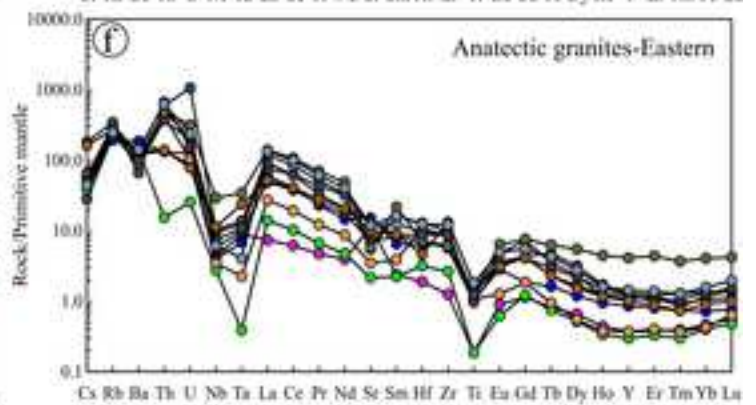
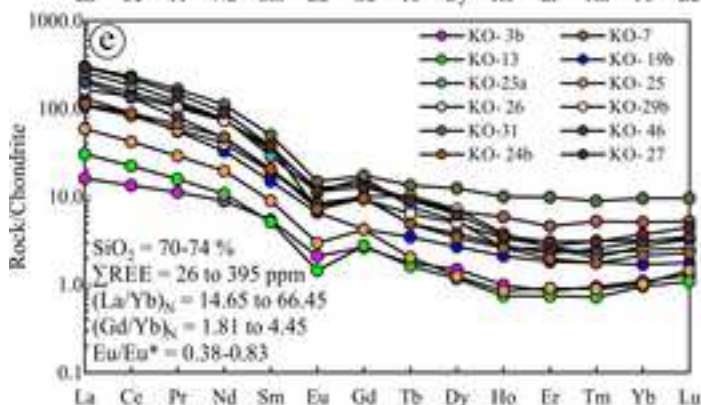
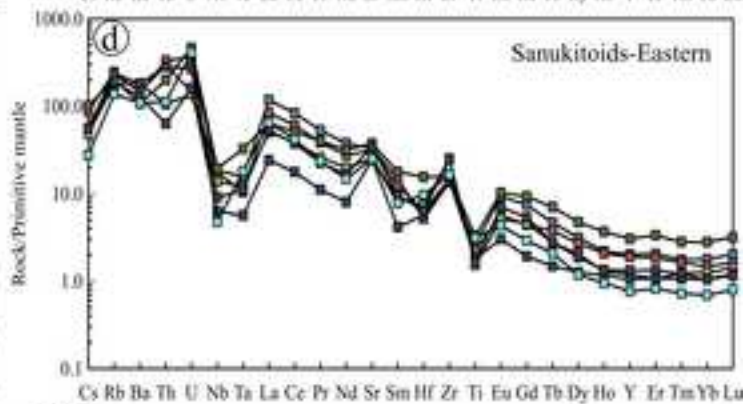
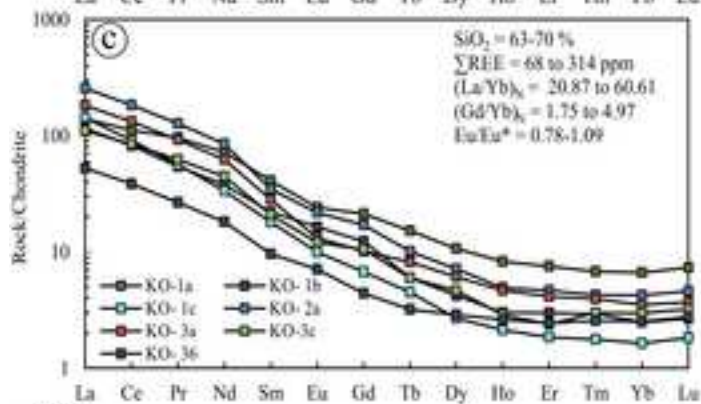
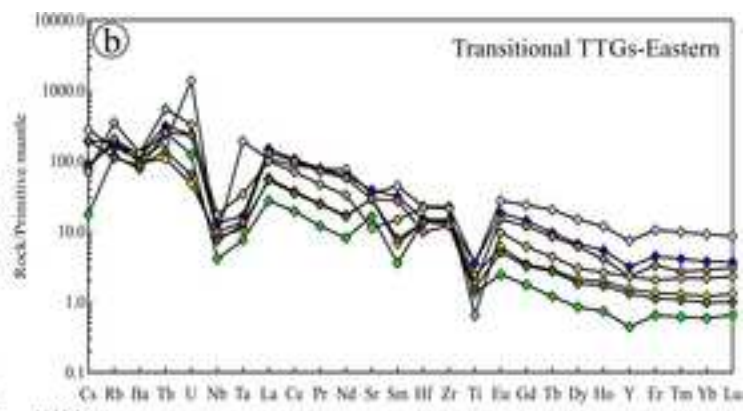
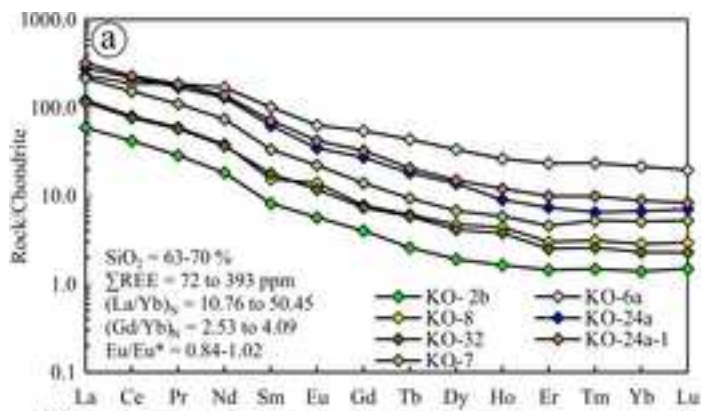


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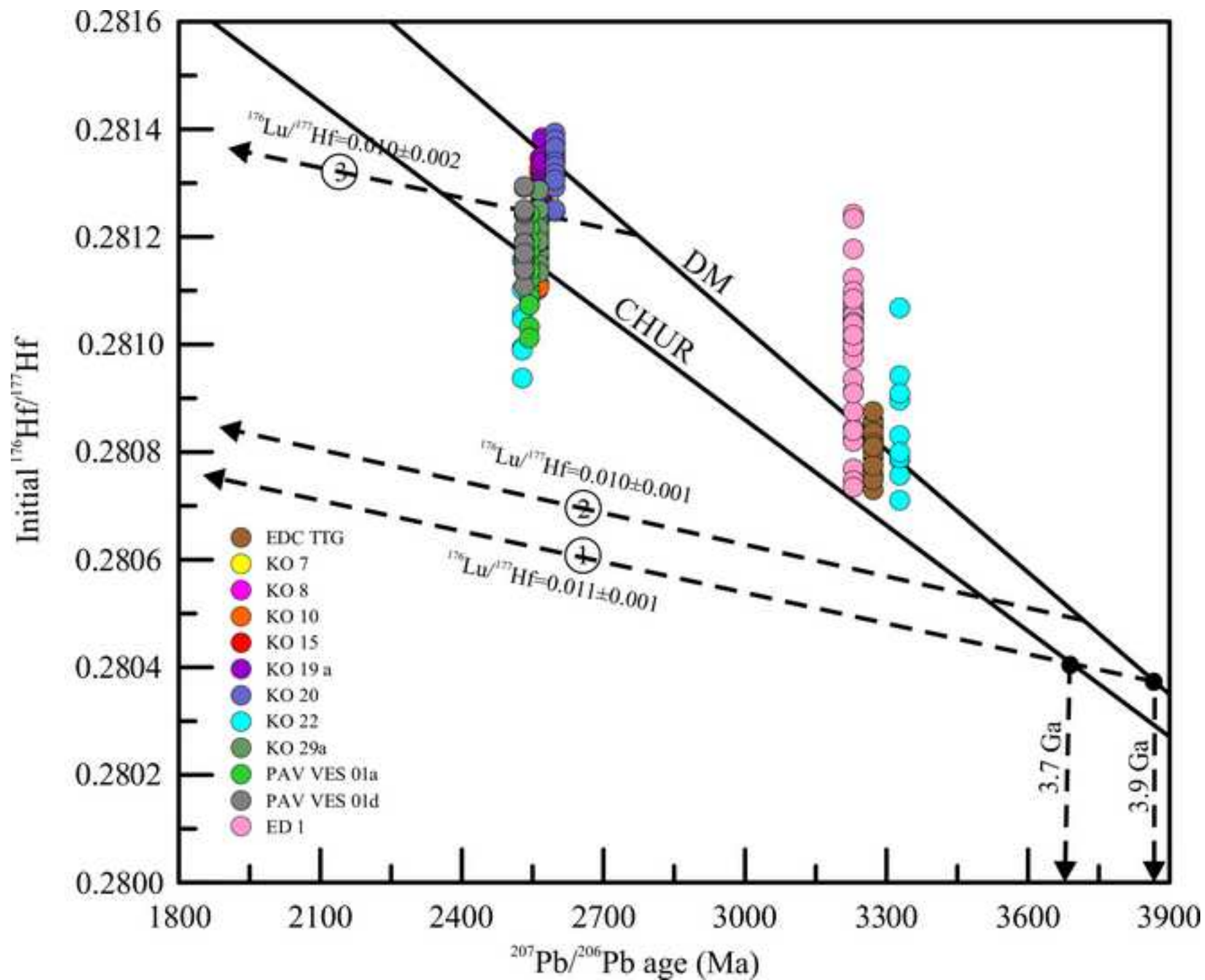


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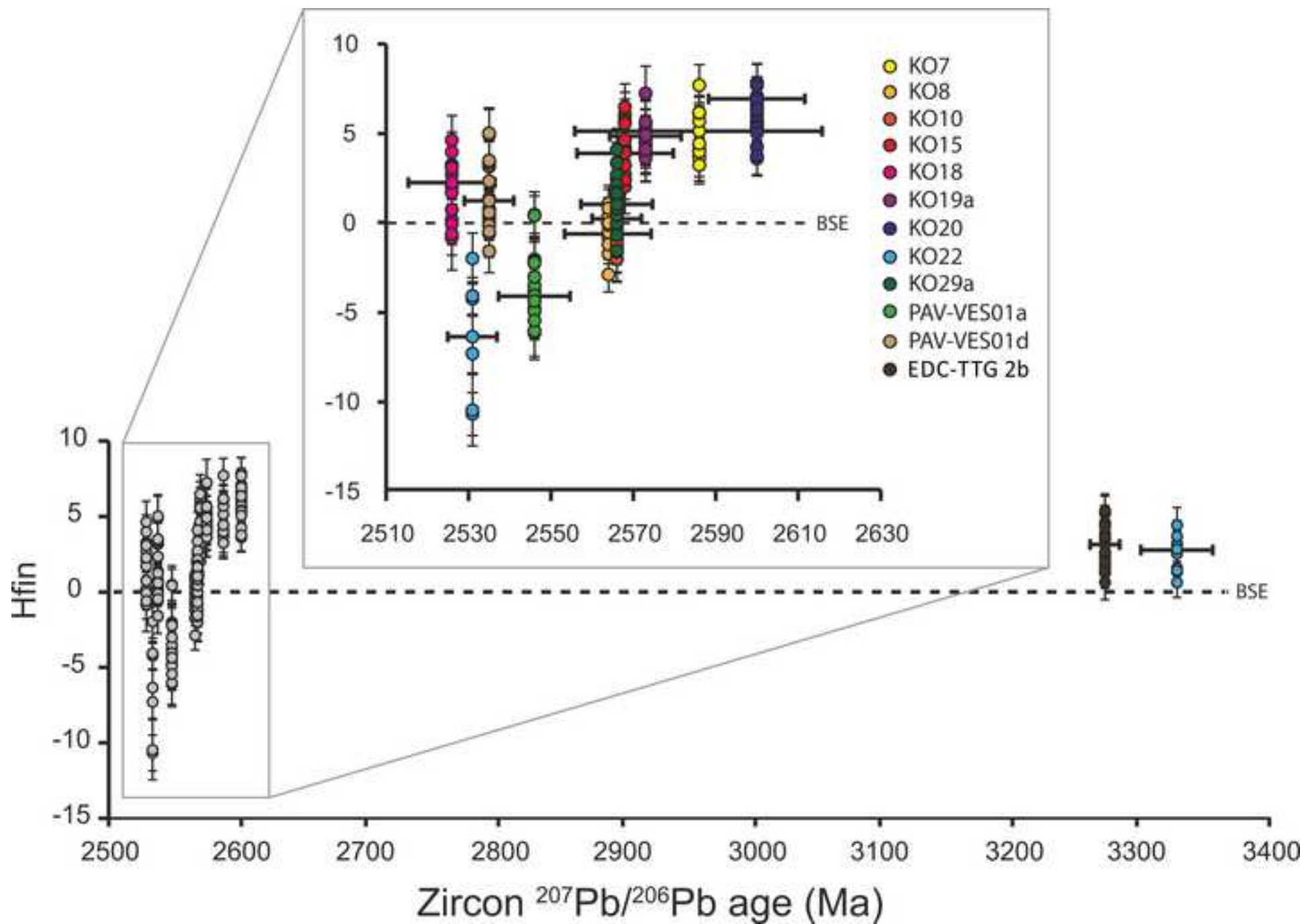
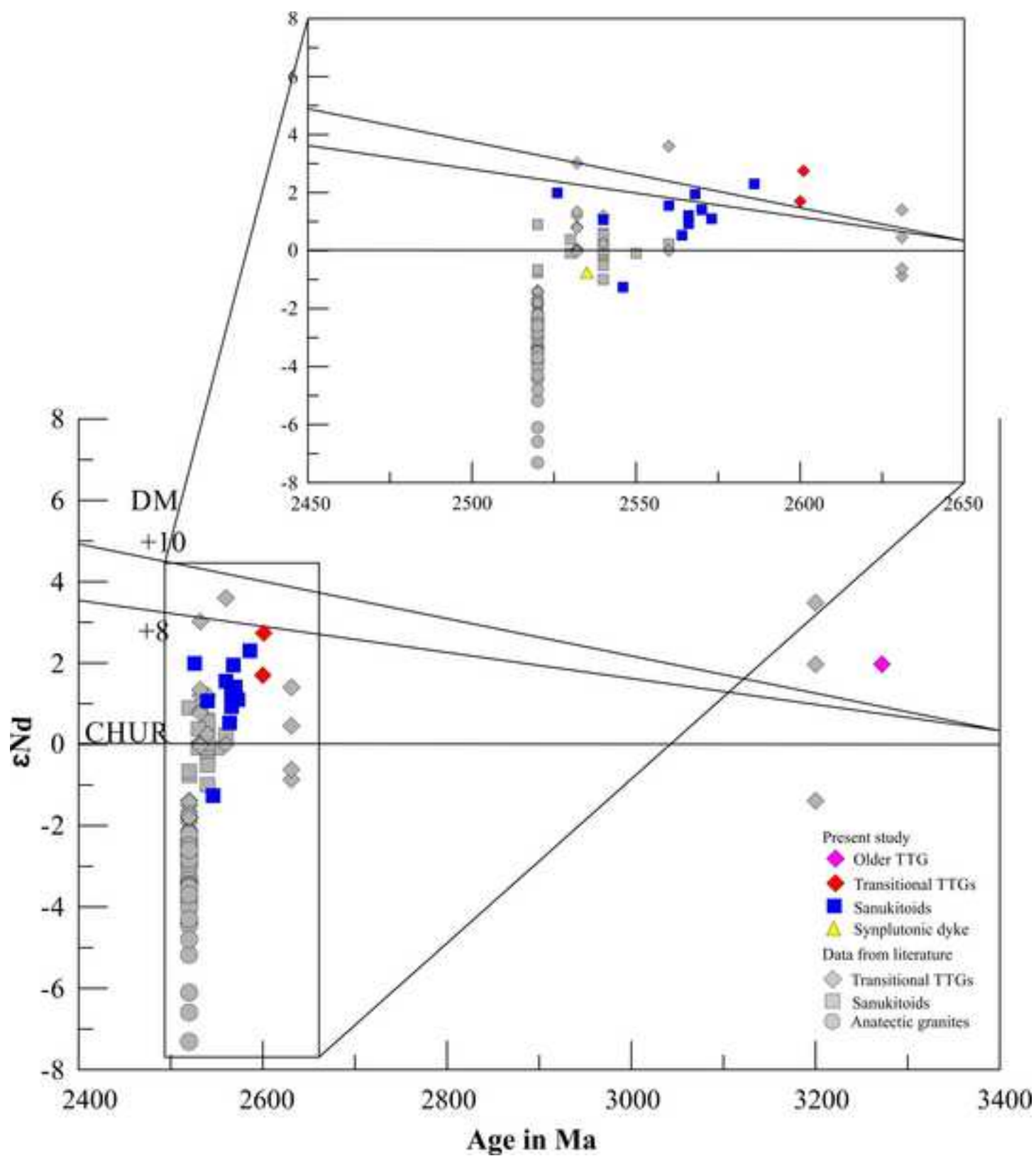


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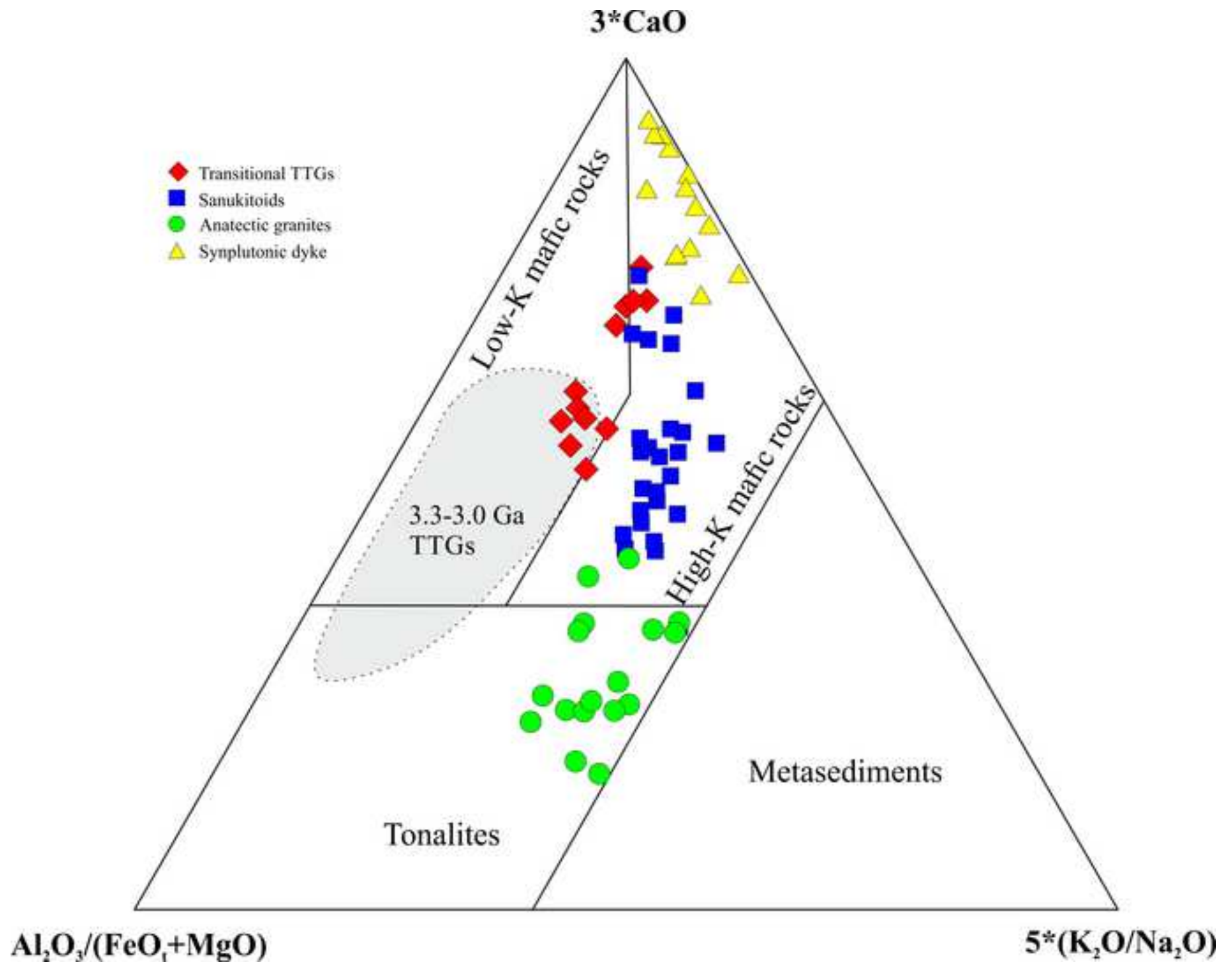


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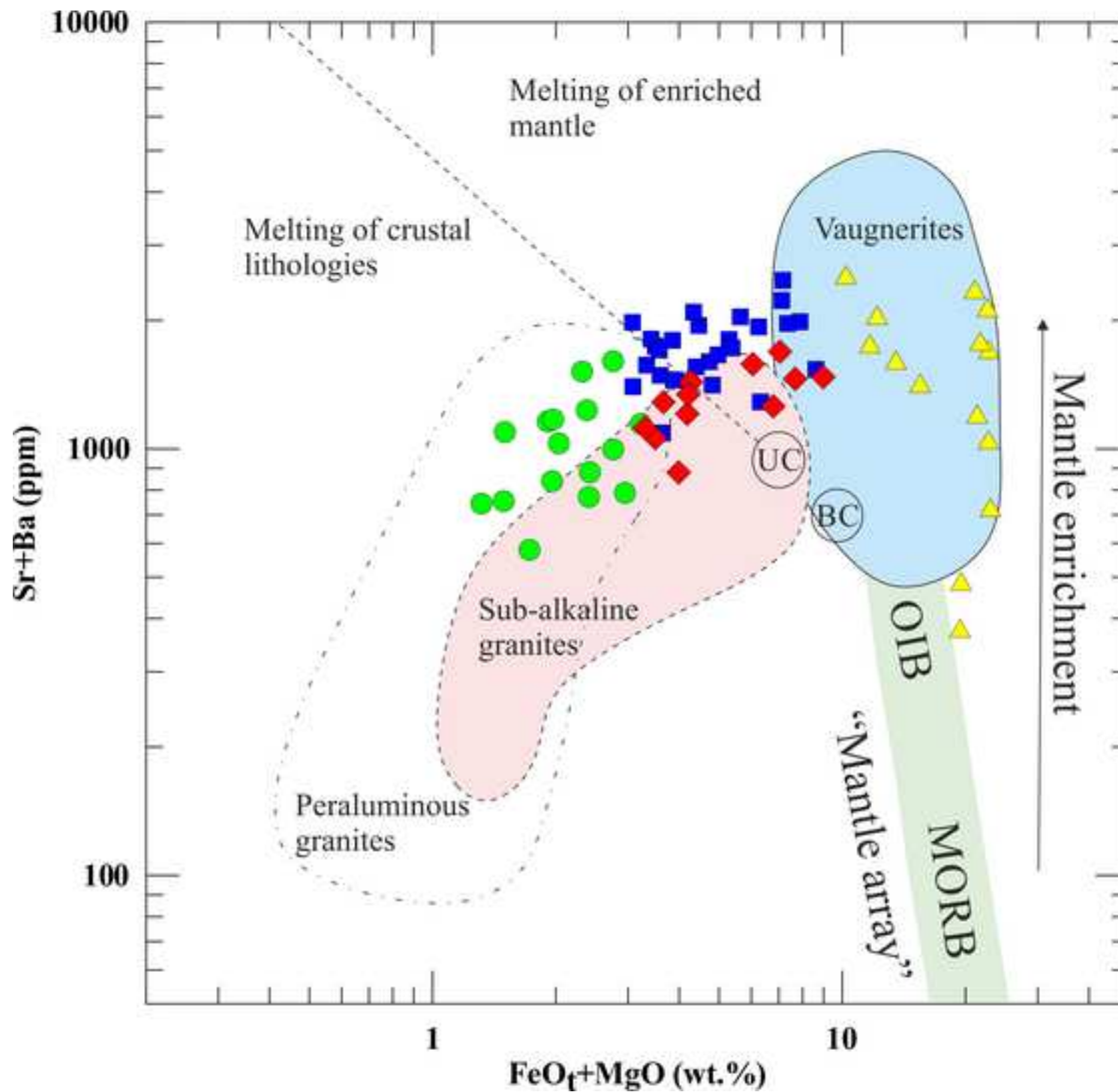


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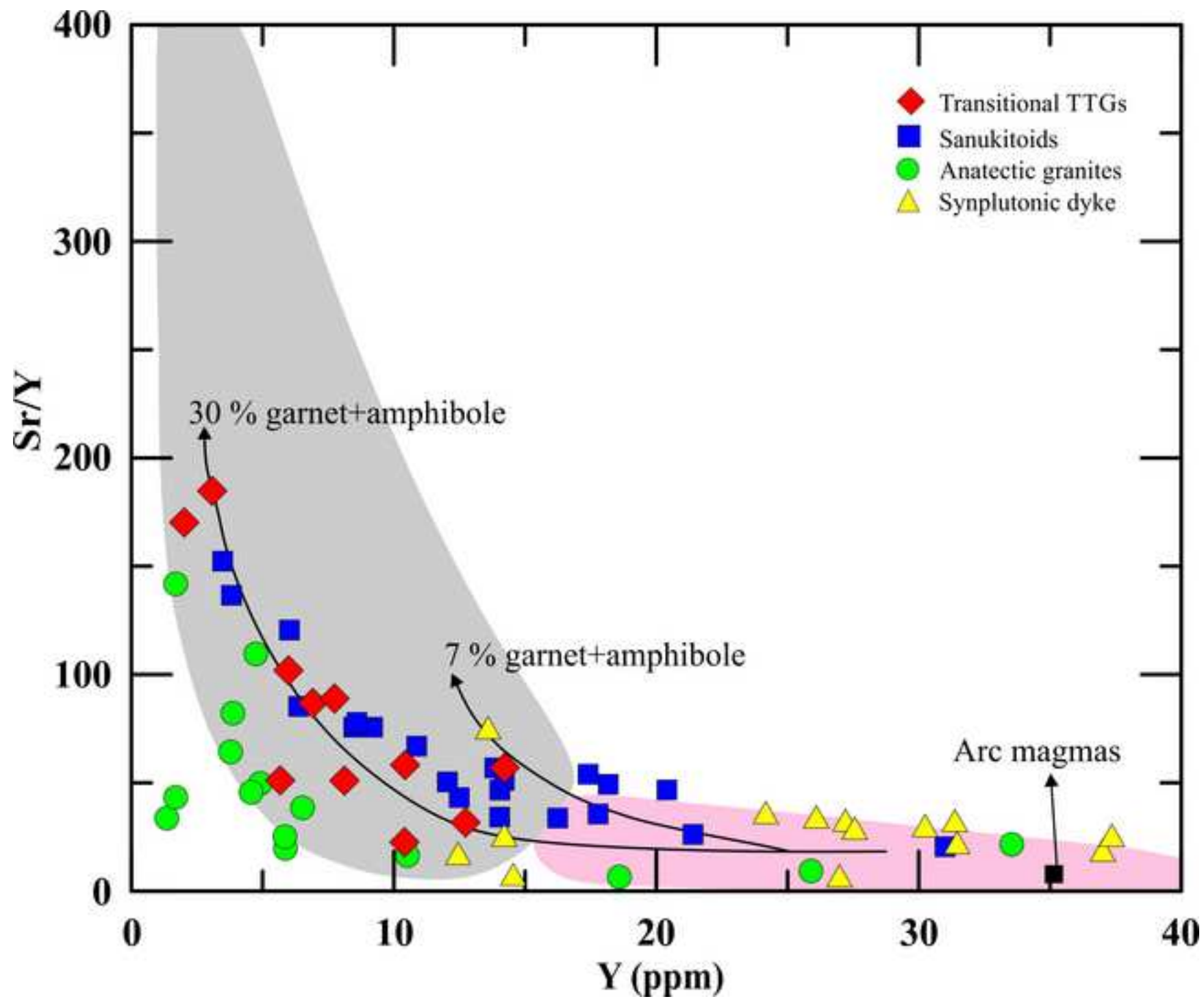


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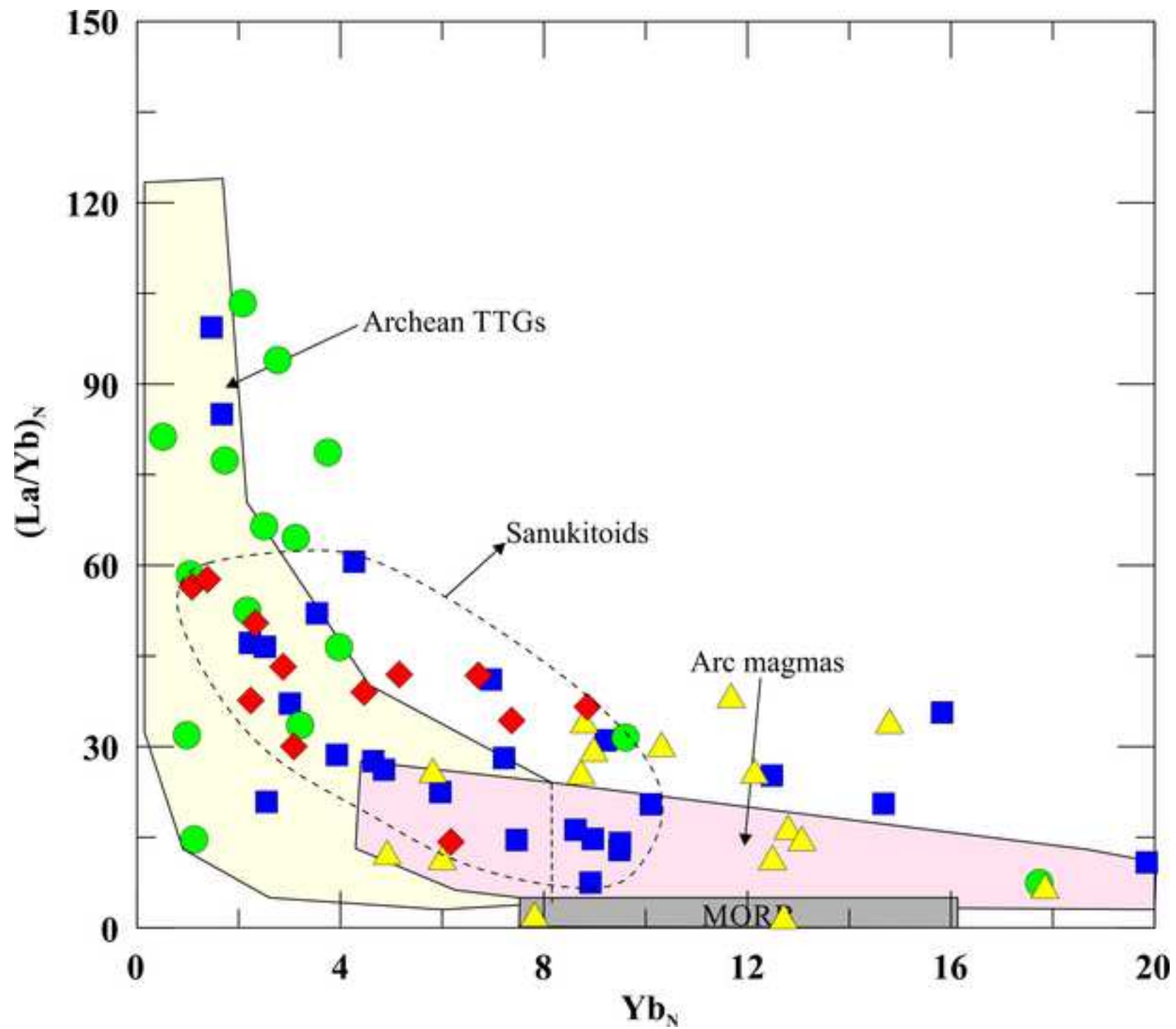
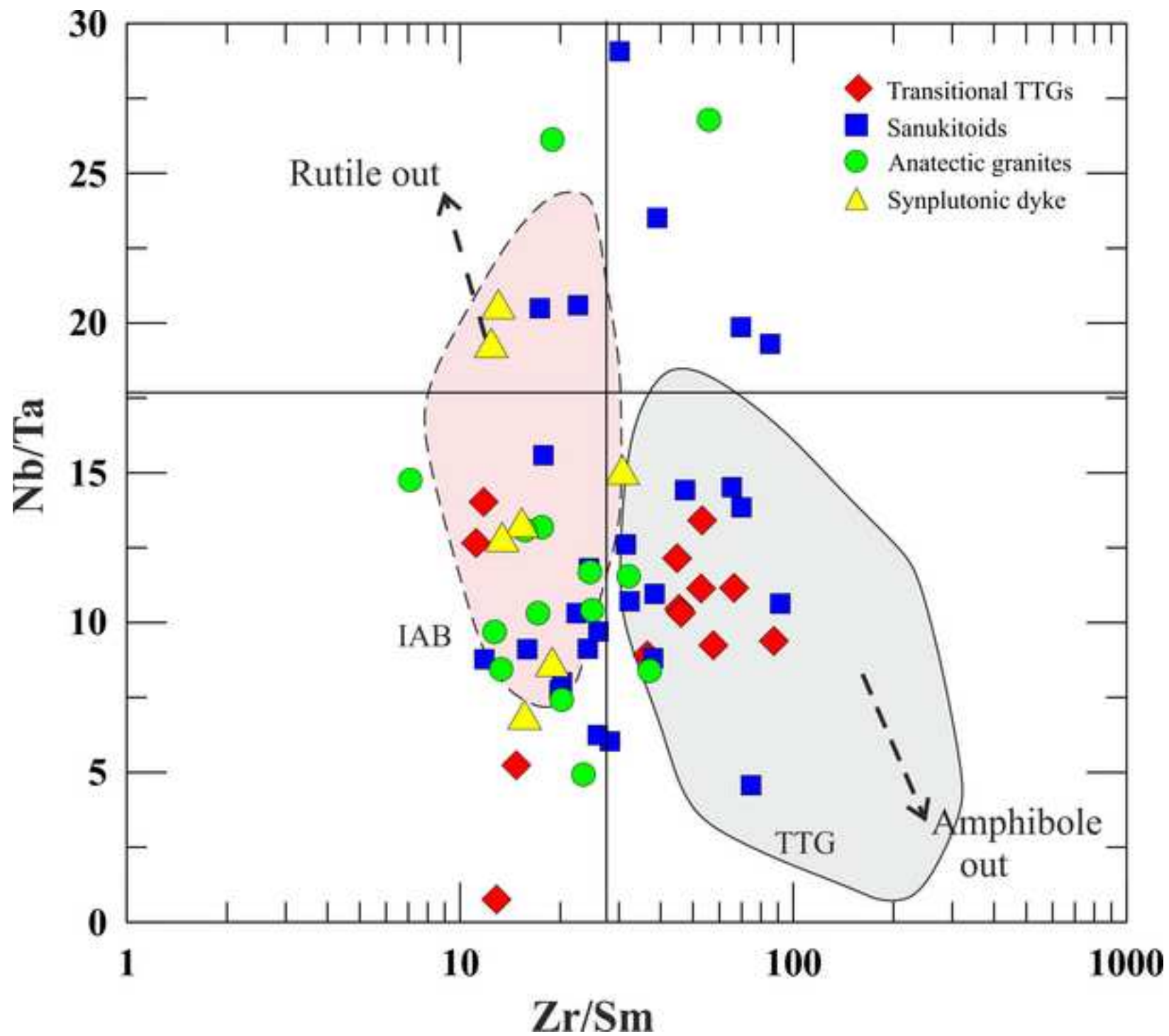


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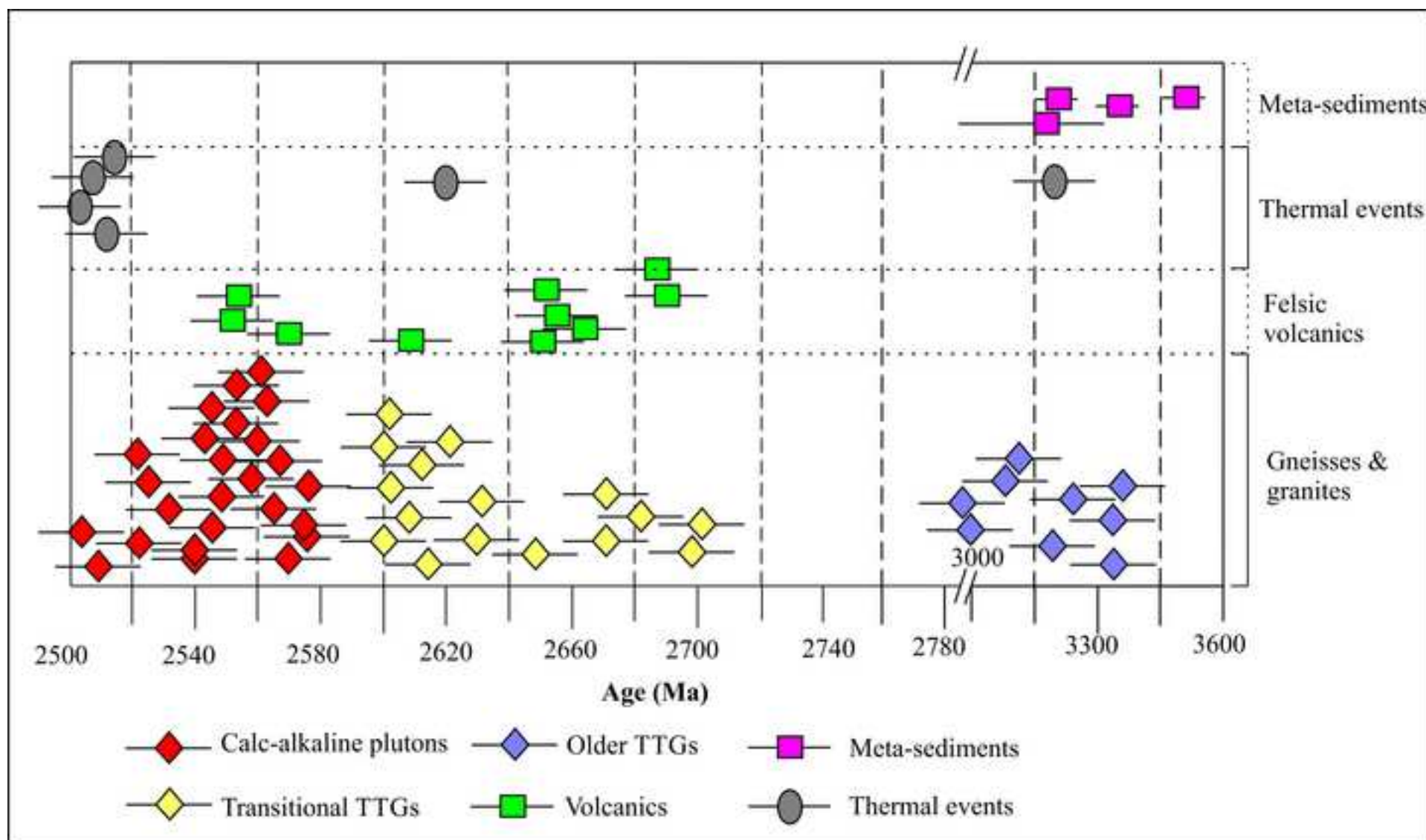
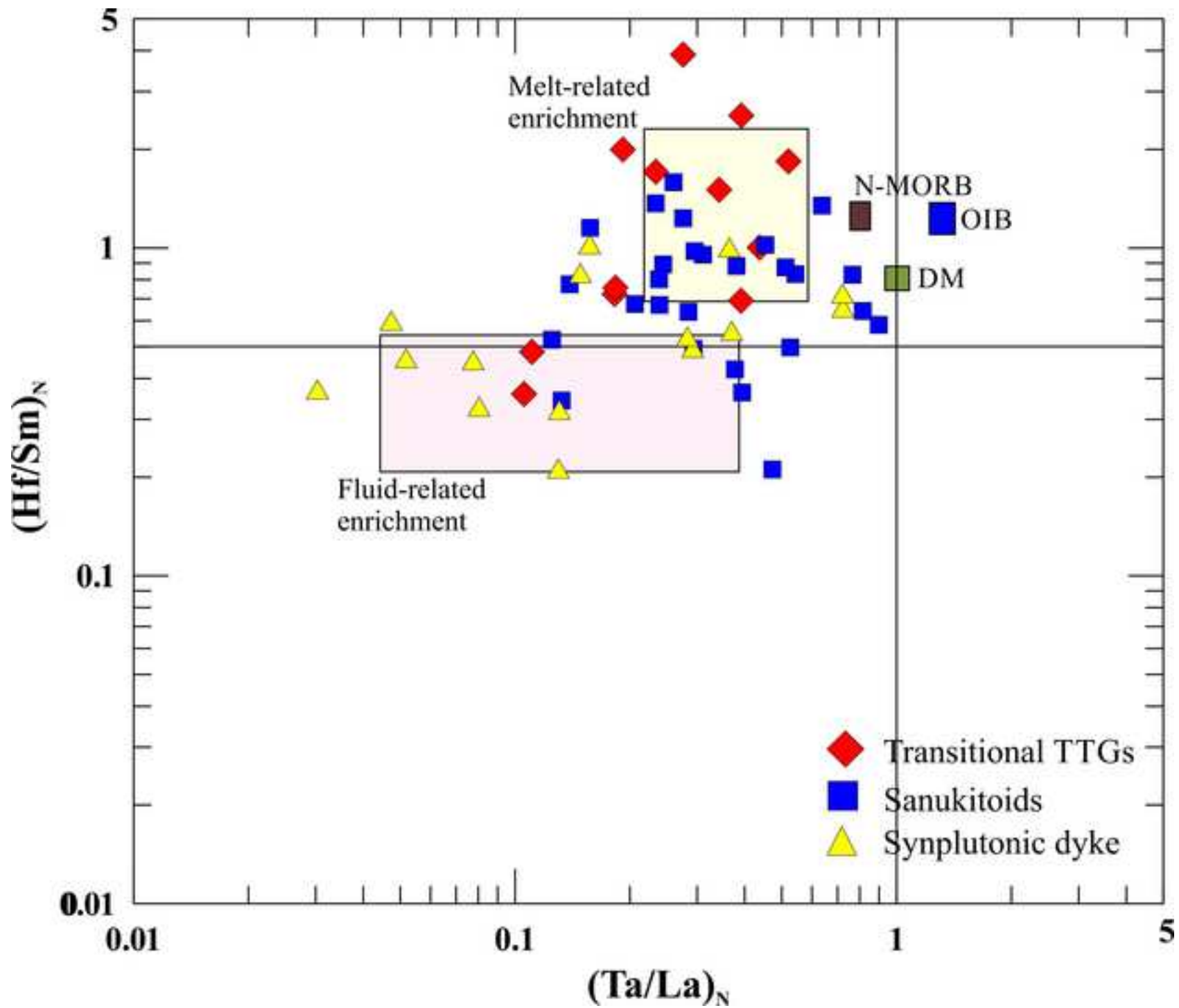


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