TECTONOTHERMAL HISTORY OF THE GRANULITES AND GNEISSES AROUND PHULBANI, ODISHA, INDIA AND ITS BEARING ON THE EVOLUTION OF THE PROTEROZOIC EASTERN GHATS BELT

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by

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DISCUSSION

Structural, petrological, fluid and geochronological data described in the previous chapters can now be combined to understand the geological evolution of the Phulbani domain. In this chapter, the evolutionary history will be discussed using different geological perspectives. These will be presented in different sections and subsections. Consequently, the tectonic buildup of this crustal domain will be highlighted in the framework of the Eastern Ghats Province and to a larger context of the supercontinent Rodinia.

10.1 Geological evolution of the Phulbani domain

Based on the geochronological data (summarized in figure 10.1), it can be suggested that the basement formation at the Phulbani domain initiated ca. 1173 Ma. This basement is now preserved as felsic augen gneiss at the northern part of the domain. Subsequent sedimentation occurred over this basement and the sediments (now represented as aluminous granulite and calc-silicate granulite) underwent an initial phase of metamorphism (M_1) which eventually culminated in UHT metamorphism (M_2 at 1000°C and 8 kbar) at ca. 987 Ma. The aluminous granulite registered growth of silica-undersaturated corundum-bearing mineral assemblage during pre-M_2 stage and subsequently produced silica-saturated spinel+quartz-bearing peak- M_2 mineral assemblage at the UHT stage. It is important to note that a biotite+sillimanite+ilmenite-bearing mineral assemblage, which occurs only as included phases within the porphyroblastic minerals of aluminous granulite, is interpreted to represent the M_1 metamorphism at amphibolite facies condition. The peak M_2 metamorphism in calc-silicate granulite is represented by scapolite+clinopyroxene+wollastonite+plagioclase-bearing mineral assemblage which stabilized at temperature >800°C. Both the above rocks were affected by isobaric cooling (M_{2R}) following the UHT metamorphism and produced sillimanite and garnet corona over spinel in aluminous granulite and garnet corona over scapolite, clinopyroxene, wollastonite and plagioclase in calc-silicate granulite. In the fine-grained charnockite gneiss, the isobaric cooling (M_{2R}) produced intergrowth of garnet and quartz around orthopyroxene and ilmenite approximately at 800°C. Cooling also produced biotite-bearing mineral assemblage in the migmatitic felsic gneiss. The age of the cooling is estimated to be ca. 968 Ma from the monazite of migmatitic felsic gneiss. The P-T-t path of the M_1-M_2 stage is characterized by nearly isobaric heating-cooling type (Fig. 10.2). Field data further suggest that the charnockite magma (precursor of the coarse-grained charnockite) was emplaced within the UHT crust and the oscillatory-zoned zircon domain yields the emplacement age as ca. 970 Ma. This rock preserves evidences of cooling related texture like intergrown garnet and quartz around orthopyroxene and thus witnessed the later stage of M_2 metamorphism (M_{2R}). The pressure of M_{2R} stage was estimated to be ~6.5 kbar using the garnet-orthopyroxene-plagioclase-quartz assemblage present in both varieties of charnockite which is lower than the pressure of peak metamorphic condition. It possibly implies minor pressure decrease (ΔP=1.5 kbar) during near-isobaric cooling (M_{2R}). Zircon grains from the migmatitic felsic gneiss further record overgrowth formation at ca. 949 Ma which possibly reflects a phase of high-grade metamorphism, but textural data do not reveal any such specific imprint on the UHT mineral assemblages. The timing of this latter phase of metamorphism is also documented in monazite grains of aluminous granulite and migmatitic felsic gneiss as ca. 939 Ma and ca. 931 Ma ages which broadly coincide with ca. 949 Ma zircon forming event. The major tectonothermal events at the Phulbani domain were complete by ca. 900 Ma as no younger group ages are reported from this domain. The Phulbani domain was subsequently transected by several ductile shear zones. Few monazite data from the shear zone rocks suggest possible timings of ductile shearing at ca. 781 Ma (N-S trending ductile shearing during $D_{4'}$) and ca. 551-531 Ma (E-W trending shearing at the RSZ during D_5). Based on the biotite-bearing mineral assemblages, it can be suggested that shearing activity occurred lower than granulite facies conditions (i.e. mid-to shallow crustal depth). This is vindicated by the fact that zircon grains from these shear zone rocks did not register any growth during these events.

All the rocks of the Phulbani domain equilibrated with a CO_2-rich fluid during the peak-UHT metamorphism as evidenced from the fluid inclusion data. Dominance of CO_2-rich fluid throughout the evolutionary history of the granulite suite indicates sustained thermal high conditions of the lower crust. Similar CO_2-rich fluid inclusions are noted from many HT/UHT granulite terranes (Cuney et al., 2007; Tsunogae and van Reenen, 2007; Santosh et al., 2008; Huizenga and Touret, 2012) including the EGP (Mohan et al., 1997; Sarkar et al., 2003; Bose et al., 2009). Absence of H_2O-rich fluid inclusions in these terranes suggests that H_2O was lost from the rock system during progressive metamorphism either as fluid phase or partitioned into the anatectic melt at the granulite facies condition. The melt, on the other hand, must be removed from the lower crust to preserve the HT/UHT mineral assemblages (White et al., 2002). The source of CO_2 within the lower crust is debatable (Touret and Huizenga, 2011; Huizenga and Touret, 2012). CO_2 can be sourced internally (Touret, 1971; Hollister, 1988; Cesare et al., 2005; Cesare and Maineri, 1999) during metamorphism. Presence of calc-silicate rocks in Phulbani domain can be considered as a possible source of internally produced CO_2. On the other hand, CO_2 fluid may come from external sources (e.g. mantle) during crystallization of basic rocks in the lower crust (Touret, 1992; Santosh and Tsunogae, 2003). There is no such rock in the study area, but possibility of their occurrence below the present surface level cannot be ruled out. The relationship between CO_2 infiltration/flushing and HT/UHT metamorphism has been extrapolated in the scale of global carbon recycling and evolution of tectosphere (Santosh and Omori, 2008). It is speculated that the UHT rocks might represent the windows for transferring CO_2 from the mantle to the crust and eventually to the atmosphere during the supercontinent cycle (Santosh and Omori, 2008; Touret and Huizenga, 2012).

Apart from CO_2-rich fluid, textures related to alkali-bearing fluid are also present within the rocks of the Phulbani domain. This is evident from the Th-rich overgrowth on the monazite in the migmatitic felsic gneiss of RSZ and presence of myrmekite-like texture and pegmatoidal rock at the contacts of calc-silicate granulite and coarse-grained charnockite. In the latter, spatial distribution of the textures suggests that coarse-grained charnockite could be the source of the alkali-bearing fluid which reacted with calc-silicate rock soon after the charnockite intrusion. The details of the reactions and related textures concerning the presence of the alkali-rich fluid have been discussed in the later section (section 10.1.4).

10.1.1 Aluminous granulite and the corundum problem

The mineral assemblage present in the aluminous granulite contains both corundum and quartz which is enigmatic and merits special attention. The rock shows contrasting mineral assemblages of Spl_2-Crn_1-Grt_1 and Spl_1-Sil_2-Qtz during the course of progressive metamorphism. Presence of Grt_2+Sil_3 corona over Spl_1 and Spl_2, Grt_2 around Sil_2, granules of Crn_2 within Grt_2 and Spl_1/Spl_2 could result from near-isobaric cooling. Textural evidences show that Spl_1, Ilm_2, Grt_1 and Crn_1 in both the mineral associations appear as a result of dehydration melting. Occurrence of corundum (Crn_1)-bearing and quartz-bearing assemblages in the same rock indicates preservation of two stages of progressive metamorphism and the transformation of the former mineral assemblage (Pre-M_2) to the latter (peak-M_2) is interpreted to occur during a heating dominated prograde path.

Porphyroblastic Crn_1 and Grt_1 (both having Sil_1 inclusions) are found to coexist in all the associations although their coexistence is interpreted as rare (Dharmapriya et al., 2015). This garnet + corundum association is indicative of high pressure and temperature conditions for Mg-rich bulk composition (Kelsey et al., 2006; Osanai et al., 1998; Shimpo et al., 2006; Tsunogae and Santosh, 2007). For pyroxene-free assemblages, composition of garnet provides crucial indication of the pressure of metamorphism. Shimpo et al. (2006) reported garnet + corundum bearing rocks from the Palghat-Cauvery shear zone of Southern India and demonstrated that the rocks might have suffered eclogite facies metamorphism for garnet and corundum to coexist. In a direct response, Kelsey et al. (2006) argued that the two phases may coexist well below the eclogite facies condition if the bulk composition is Fe-rich. The phase diagram modeling in the present study simply confirms the argument of Kelsey et al. (2006).

Coexistence of porphyroblastic corundum and quartz are reported in both the associations of the presently studied aluminous granulite. Notably, in all instances, these two minerals are separated with each other by sillimanite and/or garnet corona. When only sillimanite corona occurs in between corundum and quartz, the coexistence may be considered metastable. On the other hand, a double corona separating corundum and quartz suggests a non-equilibrium relationship between the latter phases. It is known that corundum and quartz never coexist stably over the entire P-T range of the crust as well as the lithospheric mantle (Berman, 1988; Shulters and Bohlen, 1989; Harlov and Milke, 2002; Harlov et al., 2008; Kihle et al., 2010; Kato et al., 2011) which is also confirmed in the present study. Interestingly, Shaw and Arima (1998) reported mineral assemblages similar to the present one from the Raygada area of the EGP and interpreted that the two phases coexisted stably in aluminous granulite due to ultrahigh pressure (UHP) metamorphic conditions. The argument of Shaw and Arima (1998) was based on the calculated equilibrium of Anovitz et al. (1993) contradicting the internally consistent thermodynamic data (Berman, 1988; Gottschalk, 1997; Holland and Powell, 1998) that do not allow corundum and quartz to become a stable mineral assemblage (Harlov and Milke, 2002; Harlov et al., 2008). Later studies found the interpretation of Shaw and Arima (1998) problematic as no other exclusive evidences of UHP metamorphism were found in their rock (Dasgupta and Sengupta, 2003).

Despite this, corundum and quartz are described from many regional granulite terranes and the two minerals coexist metastably in variety of P-T conditions (Mouri et al., 2003; Kihle et al., 2010; Kato et al., 2011). Harlov et al. (2008) experimentally showed that the two minerals can coexist metastably over long geologic time and the occurrence of any other mineral (particularly sillimanite) at the interface of these two will be augmented by the activity of an aqueous fluid (H_2O and/or brine). In the present case, the fluid chemistry at the peak metamorphic and subsequent cooling stage was dominated by CO_2, which has very little role in diffusive transfer of elements even at UHT condition (Harlov et al., 2008). As a result, the metastabilty between corundum and quartz could have lasted long. Raman spectral bands from few fluid inclusions, on the other hand, imply minor amount of H_2O species. Although bulk of the partial melt have left the system, the minute fraction remained in the system (such as in association A) would eventually crystallize and release H_2O. This H_2O could have been instrumental in bringing out the necessary kinetic favor to produce sillimanite at the interface of corundum and quartz. Absence of H_2O fluid species in the inclusions can be caused by its selective leakage during crustal upliftment as described in many terranes (Hollister, 1988, 1990; Touret, 2001).

Figures 5.2b and 10.2 show that corundum and quartz appeared in two different temperature windows during the prograde evolution. Two back to back reactions, i.e. Grt_1 + Crn_1 = Spl_2 + Sil_2 followed by Grt_1 + Sil_2 = Spl_2 + Qtz, are considered responsible for a shift in Si-saturation of the aluminous granulite at the peak UHT condition from a pre-peak Si-undersaturated corundum-bearing assemblage. Shulters and Bohlen (1989) experimentally constrained the reaction garnet + corundum = spinel + sillimanite which has a moderate dP/dT slope of 21 bar/°C and, is sensitive to change in both pressure and temperature. The reversal of this reaction can produce garnet + corundum from spinel + sillimanite as reported from the Highland Complex, Sri Lanka (Dharmapriya et al., 2015). For the latter case, the garnet + corundum assemblage was produced as a result of isobaric cooling at a pressure of ~9.5 kbar. In the present study, however, porphyroblastic garnet (Grt_1) and corundum (Crn_1) are significantly large in size and were produced during the prograde heating stage. However, granular and vermicular corundum within spinel and coronal garnet possibly formed during the post-UHT cooling stage. Formation of these fine corundum grains indicates that microdomainal Si-undersaturation too occurred during cooling possibly due to sluggish nature of element mobility. The calculated phase diagrams (Figs. 5.2b) show that although corundum is reported from many UHT terranes (Horrocks 1983; Osanai 1998; Sengupta et al., 1999; Ouzegane et al., 2003), it might not be stable at the peak UHT condition. Rather, it stabilizes during prograde evolution of a Si- undersaturated Fe-rich pelitic bulk composition.

Si-saturation at the UHT condition from pre-UHT corundum-bearing Si-undersaturated assemblages was previously reported (Goscombe, 1992; Kelsey et al., 2005). Goscombe (1992) used the petrogenetic grid of Hensen (1987) to show that both prograde and retrograde variety of corundum were stable without any definitive evidence of peak metamorphic corundum due to the existence of a corundum-absent invariant point at high temperature. From the relationship between the calculated Si-saturated and Si-undersaturated FMAS equilibria, Kelsey et al. (2005) noted that quartz- and corundum-bearing equilibria are linked via a set of quartz + corundum absent univariant reactions emanating from both the corundum- and quartz-absent invariant points. This is also evident in a system close to FMAS, where quartz can be produced with spinel (or sapphirine in more Mg-rich compositions) if the tie lines involving the assemblages garnet+sillimanite and orthopyroxene+sillimanite are crossed resulting in transformation to a quartz-present assemblage from a number of quartz-absent assemblages in course of prograde metamorphism (Bose et al., 2000). McDade and Harley (2001) encountered a similar situation in their theoretical KFMASH petrogenetic grid where quartz-absent invariant points are found to be metastable at higher temperature side of the grid and quartz became a part of the stable mineral assemblages although they did not consider corundum in their grid. In a comprehensive summary, Kelsey et al. (2005) showed that in many natural occurrence data, corundum appeared in the early stage of high-temperature metamorphism and eventually overprinted by spinel- and sapphirine-bearing assemblages during further heating. These authors also noted that Si-undersaturated mineral assemblages often switch stability from low-Si to high-Si fields during prograde stage and vice versa to exhibit multitude of reaction textures involving corundum, sapphirine and spinel. Some rocks also belong to a “silica-influenced transition zone” which lies between the corundum-absent and quartz-absent invariant points in the petrogenetic grids. All of the above studies suggest that a Si-understaurated rock may transform to Si-saturated one and produce quartz-bearing mineral assemblages in course of progressive metamorphism even without any addition of externally derived Si.

Kelsey et al. (2005) discussed the implications of the presence of quartz in Si-undersaturated bulk and identified four possible reasons: (1) quartz occurs either as inclusions and totally removed from the matrix during prograde reaction, (2) quartz appears due to progressive change of the assemblage from quartz-absent fields to quartz-present fields, (3) quartz appears as late phase with feldspar as a part of crystallized melt that can enter the rock during post-metamorphic stage, and (4) quartz may be retained in microdomain-scale due to sluggish reaction kinetics in an otherwise Si-undersaturated rock. Out of these, we consider the second reason as the primary factor for occurrence of quartz in the corundum-bearing aluminous granulite. The reason behind shifting from quartz-absent, corundum-bearing assemblages to corundum-absent and quartz-bearing assemblage is the progressive reaction during heating of the lower crust through a temperature window of ~850 to ~950°C (Fig. 5.2b). The occurrence of thick quartz and K-feldspar leucosome layers surrounding the aluminous granulite must have provided additional quartz surrounding the Si-undersaturated domains which justifies the reason (3) of Kelsey et al. (2005).

10.1.2 Timing of UHT metamorphism and charnockite magmatism in Phulbani domain

As garnet was produced at the peak UHT metamorphic (M_2) stage in aluminous granulite, data from the monazite inclusions are expected to record the maximum age of the UHT metamorphism as garnet grew sometime after monazite included within it. Therefore, the group age of 987±6 Ma from the monazite grains of aluminous granulite correspond to the maximum age of UHT metamorphism (M_2) at Phulbani domain. Although no accurate age of UHT metamorphism could be ascertained from zircon data of the aluminous granulite, monazite data provide the important information.

Th-rich, Y-depleted oscillatory zoned monazite cores and few Y-rich monazite patches in the aluminous granulite and the migmatitic felsic gneiss further yield ca. 966 Ma and 968 Ma ages. Presence of oscillatory zoning at the core (Fig. 9.6q) suggests crystallization of monazite from a melt phase. Depletion of Y at the core implies monazite crystallization from melt postdated garnet crystallization (Foster et al., 2000; Williams et al., 2007). On the other hand, Y-rich patches in the same monazite could result from the breakdown of REE-bearing minerals like xenotime and/or garnet which are considered to be major repositories of Y (Zhu and O'Nions 1999; Williams et al., 2007). In absence of xenotime in both the migmatitic felsic gneiss and the aluminous granulite, Y must have been sourced from garnet only. Pyle and Spear (2003) observed discontinuous rims of monazite with higher Y content (similar to the Y-rich patches of the present study) compared to interior monazite domains of migmatitic gneiss and they interpreted the formation of Y-rich monazite by garnet breakdown during melting. Such late overgrowth on monazite is a natural consequence of cooling when Y diffuses out from garnet in anatectic migmatites and interestingly, monazite grown under such conditions may display only minor age variability across relatively sharp compositional boundaries (Spear and Pyle, 2010). Thus, it is apparent that ca. 966 and 968 Ma monazite ages from the aluminous granulite and the migmatitic felsic gneiss represent the age of cooling (M_{2R} in the present study) following the UHT metamorphism. These latter ages additionally help constraining the absolute age of the peak UHT metamorphism of the present study which must have occurred before melt crystallization but after ca. 987 Ma (maximum age of peak metamorphism constrained from the included monazite).

Interestingly, Upadhyay et al. (2009) studied one sample from the Phulbani domain from which they interpreted the age of UHT metamorphism at ca. 1.2 Ga based on zircon U-Pb data.

They noted that the ca. 1.2 Ga ages were obtained mainly from the reequilibrated oscillatory-zoned relict cores and overgrowths of zircon. It is, therefore, quite likely that the ca. 1.20 Ga zircon age could be inherited given the internal morphology of the zircon grains. Moreover, these authors did not elaborate the tectonic setting of the UHT metamorphism that they interpreted to occur at ca. 1.2 Ga.

Coarse-grained charnockite contains fragments of aluminous granulite and calc-silicate granulite which indicate that intrusion of magmatic protolith of this charnockitic rock after the peak metamorphism. Additionally, it is also important to note that the presence of the isobaric cooling related reaction texture in this rock suggests that charnockite magma must have intruded the UHT suite and crystallized just after the peak metamorphism and subsequently cooled along with aluminous granulite and calc-silicate granulite. Group ages of 970±6 Ma and ca. 973 Ma from oscillatory-zoned zircon grains of two samples suggest crystallization ages for charnockite magma. Based on these ages, it can be argued that the Phulbani domain was affected by charnockite magmatism at ca. 970 Ma and this provide minimum constraint on the timing of the UHT metamorphism. These ages however, fall well within the uncertainty limit (~±20 Ma following Schmitt and Vasquez, 2017) of the UHT metamorphism which suggests that UHT metamorphism and charnockite magmatism in the Phulbani domain was closely spaced in time and possibly resulted in a single thermal regime. It is important to note that the absence of S_2/S_3 gneissic fabric in the coarse-grained charnockite and presence of the same fabric in the aluminous granulite and the calc-silicate granulite clearly suggests that there is a definite time-gap between UHT metamorphism and charnockite magmatism. This time-gap cannot be established by geochronological data as the ages appear to be same within uncertainty when data accuracy is considered. The crystallization age of the coarse-grained charnockite of the present study (ca. 970 Ma) is well within the uncertainty limit of 985±5 Ma U-Pb zircon age of charnockite magmatism reported by Paul et al. (1990) from an adjacent locality within the Phulbani domain. Additionally, the presently obtained $^{207}Pb/^{206}Pb$ dates in the range of ca. 912−960 Ma from the bright-CL zircon overgrowth represents the broad timing of metamorphism that occurred subsequent to charnockite emplacement.

High-grade metamorphism and associated charnockite magmatism is well known from the Mawson coast, Northern Prince Charles Mountains (NPCM), Kemp land, McRobertson land of east Antarctica which are interpreted to evolve with the Eastern Ghats Province during Proterozoic Rayner-Eastern Ghats (R-EG) orogeny (Morrissey et al., 2015, 2016). Charnockite magmatism in such places is interpreted to be the result of high-temperature intracrustal melting of dry and thickened granulitic crust (Young et al., 1997). Zhao et al. (1997) argued that the ca. 1000 Ma high-grade metamorphic zone of the NPCM is intruded by charnockite plutons which were mainly derived from pre-existing subduction related crustal sources and geochemically similar to the I-type granitoids. Geochemical characters of such rocks also point towards a collision related process. Charno-enderbitic rocks of ca. 1.76 Ga which are similar to the presently studied coarse-grained charnockite are also present in the Ongole domain, located at the southern part of the EGB and occur in close proximity of the UHT rocks at that area. Sarkar et al. (2015) suggested that these charno-enderbitic rocks have arc related geochemical characters with minor crustal contamination as revealed in the trace element signatures. These authors invoked a model wherein the charno-enderbite magma possibly formed by differentiation of basic magma that was produced due to melting of mantle wedge during accretion of Dharwar craton of India and Napier complex of east Antarctica. Based on these evidences, it can be speculated that the protolith of the coarse-grained charnockite could have been originated during arc magmatism at the time of the accretion of the R-EG and cratonic India. This protolith could be crustal, but its connection with the mantle cannot be completely ruled out in absence of geochemical signatures.

10.1.3 Tectonic evolution of the Phulbani domain

Structural data suggest that the Phulbani domain was affected by five phases of deformation (D_1-D_5) which is represented by simplified block diagrams in Fig. 10.3. The earliest deformation (D_1) produced S_1 fabric which is defined by biotite+sillimanite-bearing assemblage. This S_1 fabric is preserved only as inclusion within porphyroblastic minerals such as garnet, corundum and spinel in the aluminous granulite. The composite S_2/S_3 gneissic fabric is the earliest regionally recognizable planar fabric of the study area and possibly developed in N-S oriented compressional setting. Presence of spinel+quartz-bearing assemblage along the composite S_2/S_3 fabric suggests that this fabric must have developed at UHT condition. Preservation of the spinel+quartz bearing UHT mineral assemblage along the composite S_2/S_3 additionally indicates that D_2 and subsequent D_3 deformation possibly occurred at persisting high-grade condition during the ca. 987 Ma UHT metamorphism in the Phulbani domain. The D_4 deformation produced F_4 folds which are dominantly hook-shaped, reclined and steeply northeasterly plunging and resulted due to a NE-SW compression. Charnockite intrusion (ca. 970 Ma) took place prior to the D_4 stage. Presence of biotite-bearing assemblage along the S_4 fabric in coarse-grained charnockite additionally implies that D_4 deformation took place in amphibolite facies condition.

Mylonitic foliation (S_{4S}) in the N-S trending ductile shear zones resulted in response to the $D_{4'}$ deformation and occurs as axial planar foliation of the F_4 fold similar to the D_4

deformation. Sinistral shear sense on horizontal sections additionally indicates that the compression was along NW-SE direction during the $D_{4'}$ deformation. Steep southerly plunging stretching lineations present on the mylonitic foliation suggests exhumation along this shear zone and the eastern block appears to be uplifted. Quartz and K-feldspar microstructures from this shear zone indicate deformation in dislocation creep regime. Presence of prism<a> slip in the recrystallized quartz grans indicates that the deformation took place around 550°C (Stipp et al., 2002) but minor presence of rhomb<a> slip additionally suggests that the temperature of the deformation possibly reached much lower temperature, at places. Dating deformation is difficult because it requires minerals like zircon, monazite etc. to grow at the the time of ongoing deformation. In this regard, ca. 781 Ma group age calculated from monazite of aluminous granulite may possibly represents the age of N-S ductile shearing. This ductile shear zone could be related to the development of westerly placed Nagavalli shear zone which is interpreted to transpose high-grade fabric by folding and related shear zone formation (Saha and Karmakar, 2015). Lower density of the pseudosecondary fluid inclusions in some samples indicates that entrapment of these inclusions occurred during upliftment of the cooled lower crust through the shear zone.

Migmatitic felsic gneiss, coarse-grained charnockite, fine-grained charnockite gneiss, aluminous granulite and felsic augen gneiss having S_4 fabric became folded (F_5) in the vicinity of RSZ during D_5 deformation and produced axial planar E-W trending mylonitic foliation (S_{5S}). These folds are overturned in nature and northeasterly plunging as revealed from the orientations of the intersection lineations preserved on the limbs of F_5 folds. Mylonitic foliation (S_{5S}) present in RSZ is almost E-W trending and characterized by alternate mylonitic and ultramylonitic bands. Importantly, there are variations in the orientation of stretching lineation that could be related to development of sheath folds occurring both in macro- to mesoscopic-scale and eyed folds in the migmatitic felsic gneiss. These sheath folds possibly formed during exhumation of the high-grade rocks of the Phulbani domain. On horizontal sections of the mylonitic foliation, the orientation of S-C fabrics, rotated porphyroclasts of feldspar and garnet suggest a dextral sense of shear along the RSZ. Quartz grains of the migmatitic felsic gneiss, felsic augen gneiss, aluminous granulite and the coarse-grained charnockite located at RSZ dominantly form ribbons and CPO data of these grains indicates dominant presence of prism<a> slip with minor presence of rhomb<a> slip. Similar type of ribbons completely composed of quartz were earlier documented by Hippertt et al. (2001) in mylonitic gneisses of Rebeira orogenic belt of southern Brazil and interpreted to develop due to coalescence of scattered quartz grains by crystal plastic processes. On the basis of CPO data of the quartz grains, these authors additionally suggested that the quartz grains of the ribbons were deformed by basal<a> slip at around 700°C in a dry deformation condition. In the present case, however, quartz grains are dominantly deformed by prism <a> slip possibly in the presence of a fluid phase as revealed from monazite textures of the RSZ.

Pseudotachylite veins are conspicuous in RSZ and dominantly occur along the mylonite and ultramylonite bands. Deformed clasts of the host mylonitic rock invariably occur within these pseudotachylite veins and indicate that the development of mylonite prior to the pseudotachylite formation. Presence of the deformed microlites of garnet, ilmenite etc. suggests that the pseudotachylite veins parallel to the mylonite and ultramylonite bands became deformed subsequent to their crystallization. Deformed pseudotachylite is common in rock records (Price et al., 2012; Kirkpatrick and Rowe, 2013 and references therein) and the plastic flow in such pseudotachylite occurs due to finer grain size (Chattopadhyay et al., 2008; Ueda et al., 2008).

Another intriguing fact is that the presence of mylonitic clasts within the pseudotachylite veins occurring parallel to the mylonitic and ultramylonitic bands could have resulted due to *"self-localizing thermal runaway"* process which requires a ductile precursor of a pseudotachylyte-bearing fault (John et al., 2009). Based on these evidences, it is suggested that the pseudotachylite veins as well as the mylonitic foliation possibly developed in a single deformation event. However, the other pseudotachylite veins cross cutting the mylonitic folation at high angle are always associated with N-S oriented shear fractures and interpreted to develop at shallow crustal level and possibly much younger than the mylonitic foliation of RSZ.

Structural study of Phulbani domain thus points toward a complex tectonic evolution where UHT metamorphism occurred at deep- to mid-crustal condition (25-30 km). The most suitable setting for deep to mid-crustal UHT metamorphism is the back arc setting (Currie and Hyndman, 2005) which has been invoked to explain the UHT metamorphism of the EGP as a whole with a framework of subduction-accretion system (Dasgupta et al., 2013, 2017; Bose and Dasgupta, 2018). In such subduction-accretion system, the back arc extension is followed by compression, which in the present case is manifested by the operation of D_2-D_3 stages producing the composite S_2/S_3 fabric. This compressional event is speculated to be related to post-extensional thickening of the ultra-hot orogen (Harley, 2016). Subsequent deformation phases (D_4, D_5) occurred at relatively lower metamorphic grade as evidenced from biotite-bearing mineral assemblages and interpreted to be responsible for exhumation of the UHT metamorphosed rocks of Phulbani domain to relatively shallow crustal level.

It is well known that shear zones are pathways for metamorphic fluids. The transfer of fluid through the RSZ is evident from the chemical signatures on monazite grains e.g. development of Th-rich veins and overgrowth (Figs. 9.6j, k). Similar Th enrichment in monazite has been ascribed to infiltration of alkali-bearing fluid in monazite (Harlov et al., 2011; Kelly et al., 2012). Monazite included within garnet were not supposed to be affected by such fluid ingression but orientations of the Th-rich veins and presence of pore spaces along the veins suggest that these veins simply mimic the fracture planes which perhaps developed during brecciation of the host garnet and provided the avenue for fluid migration. Crowe et al. (2001) obtained ca. $^{40}Ar/^{39}Ar$ age from recrystallized biotite present in mylonitic augen gneiss of RSZ and based on this these authors invoked that RSZ was active during the time-frame of ca. 550-500 Ma. In this regard, spot dates within the range of ca. 551-535 Ma obtained from matrix monazite of felsic augen gneiss of RSZ could be correlated to ca. 550-500 Ma time-frame of Crowe et al. (2001) and possibly suggest activity along this shear zone. Such fluid infiltration could also be responsible for partial resetting of monazite age (ca. 831 Ma) in the felsic augen gneiss.

Interestingly, the Akul fault zone and the Riamol shear zone are located at the north of RSZ and demarcate the boundary of the Rengali Province and EGP. These are interpreted to develop during ca. 521-498 Ma (Ghosh et al., 2016). Considering the similarities in the spread of ages, it can be speculated that RSZ developed along with the Akul fault zone and the Riamol shear zone as a composite shear zone system during final amalgamation of India and East Antarctica as a part of Gondwana supercontinent.

10.1.4 Metasomatism related to charnockite emplacement

Emplacement of charnockite within the calc-silicate granulite produced myrmekite-like intergrowth and patchy Ca-rich zones in the plagioclase grains of coarse-grained charnockite. Additionally, myrmekite and albitic film in the pegmatoidal rock resulted due to involvement of

Na and K from coarse-grained charnockite. These two processes appear to be simultaneous where Ca was transferred from the calc-silicate granulite to the coarse-grained charnockite in exchange of Na and K of the former. K-feldspar lamellae within plagioclase of coarse-grained charnockite possibly developed by exsolution during cooling and BaO was partitioned into the K-feldspar during this process. Some amount of Na was also incorporated into the K-feldspar to develop Na-poor, Ca-rich zones around the lamellae. Textural characters thus indicate that the element transfer possibly occurred through a fluid phase. High density CO_2 was found in the coarse-grained charnockite as fluid inclusions in quartz and feldspar but element mobility of CO_2-rich fluid is interpreted to be low (Watson and Brenan, 1987). Therefore, it can be speculated that a fluid phase with higher element mobility must have been present during the metasomatism. Such fluid phase should have high CO_2 and low H_2O components as the metasomatism occurred at high-temperature condition during charnockite intrusion. Experimental data (Newton and Manning, 2010; Manning and Aranovich, 2014) indicate that such a fluid phase could be hypersaline brine with alkaline affinity which is thought to be responsible for mineral solubility and associated element mobility to greater extent. Fluid inclusion evidences from many localities suggest that saline fluids of the lower crust and upper mantle may coexist across a miscibility gap with CO_2-rich fluids (e.g. Touret 1985; Gibert et al. 1998).

The possible source of the hypersaline brine in the present case could be the coarse-grained charnockite as deep-derived magmas are sometimes considered to carry hypersaline fluid highly charged with CO_2. Such hypersaline saline fluid was not found during fluid inclusion analyses possibly due to its greater mobility (Touret and Huizenga, 2011) but presence of halite crystals in coarse-grained charnockite suggests that a component of hypersaline fluid was definitely present in this rock.

The myrmekite front of the metasomatic rock appears to be porous which may form due to differences in solubility and molar volumes between the product and the reactant phases in aqueous alkali halide solution (Putnis, 2009 and references therein). Widespread occurrence of myrmekite with pores indicates extensive metasomatism in presence of possible brine solution (Touret and Huizenga, 2011). Moreover, K-feldspar micro-veins in the migmatitic felsic gneiss can also be formed due to the effect of supercritical brine as documented from several high-grade terranes (Franz and Harlov, 1998; Harlov et al., 1998; Harlov and Wirth, 2000).

10.2 Emplacement of granite: a part of the basement of the Eastern Ghats Province?

The ca. 1173 Ma age from the oscillatory zoned zircon domains of the felsic augen gneiss within the Ranipathar shear zone is interpreted to represent the crystallization age of the granitic protolith. Upadhyay et al. (2009) reported ca. 1.21 Ga and ca. 1.20 Ga ages respectively from monazite and zircon grains of an aluminous granulite sample of the Phulbani domain and presented these as the age of early UHT metamorphism in this domain. Aftalion et al. (1988) reported upper intercept age of 1159 +59/-30 Ma from the zircon grain of augen gneiss of the adjacent Angul domain which they assigned as the crystallization age of a part of the basement. Information of these isolated dates and ages from different localities of the northern part of the EGP merges towards an important issue regarding the timing of the basement formation. It is possible that the granitic rock (now felsic augen gneiss) from the Ranipathar shear zone was a part of the basement of the Phulbani domain. Presence of 1190±18 Ma spot dates in the xenocrystic zircon cores of the fine-grained charnockite gneiss could also be correlated with the ca. 1173 Ma crystallization age of the protolith of the augen gneiss mentioned here. Recently, Nanda et al. (2018) reported ca. 1500 Ma age from magmatic zircons of sheared granite gneisses of Koraput alkaline complex which they interpreted as part of the basement of the EGP. Bose et al. (2011) also documented similar age from inherited magmatic zircon domains of peraluminous granitoid of Vishakhapatnam domain of the EGP and interpreted this as the age of the basement to the metasedimentary rocks. Therefore, it can be assumed that the basement of the EGP could be made up of granitic rocks showing variable emplacement ages. The exact nature of such multiple magmatisms in the basement is speculative at this stage in absence of more data.

It is interesting to note that except a few spots of ca. 1038 Ma and 977 Ma dates, zircon grains of this basement gneiss did not record the discrete imprints of the UHT metamorphism or subsequent events. Some spot dates (e.g. ca.1038 Ma) could represent the mixing of age domains. Therefore, it can be speculated that the ca. 1173 Ma zircon was reworked by a younger metamorphic event. Absence of discrete zircon growth during that event could result from several factors as invoked by Kelsey et al., 2008.

10.3 Phulbani and Visakhaptanam domains in the framework of the EGP and its connection with Rodinia

EGB is subdivided into several crustal provinces and domains (tectonothermal histories of these provinces and domains are summarized in figure 10.4) of which the EGP occupy a major portion. The Phulbani and the Visakhapatnam domains which belong to the EGP are separated by the Nagavalli-Vamasdhara shear zone (Dobmeier and Raith, 2003; Chetty et al., 2003), but their petrological histories are quite similar. Aluminous granulites from both these domains show cooling dominated P-T evolution following UHT conditions (~1000°C) at deep crustal level (~8 kbar) (Dasgupta et al., 1995; Korhonen et al., 2013b; Dasgupta et al., 2017 and references therein for Visakhapatnam domain; this study for Phulbani domain).

In addition to the ca. 987 Ma monazite age from the aluminous granulite, zircon grains yield spot dates of 1000±9 Ma, 1018±8 Ma, 1026±10 Ma (aluminous granulite), 1012±8 Ma and 1024±6 Ma (fine-grained charnockite gneiss). These spot dates match with the ca. 1030–990 Ma age of UHT metamorphism from the adjacent Visakhapatnam domain (Bose et al., 2011; Das et al., 2011). Field relationship between the coarse-grained charnockite and the UHT-metamorphosed granulites of the Phulbani domain (this study) also appears to be similar to that of the adjacent Visakhapatnam domain (Korhonen et al., 2013b). The charnockite magmatism event in the former (ca. 970 Ma, this study) broadly coincides with regionally extensive charnockite-enderbite magmatism at ca. 980 Ma in the latter domain (Korhonen et al., 2013b). It implies that the Phulbani and the adjacent Visakhapatnam domain were possibly contiguous terranes within the broad framework of the EGP (Figs. 10.5, 10.6).

A similar tectonothermal history of the Rayner Province can be of interest here. A phase of charnockite plutonism (Mawson charnockite ca. 990–960 Ma) occurred in the MacRobertson Land (Halpin et al., 2007a) and the magma was found to cross-cut the high-grade gneissic banding (Halpin et al., 2007b) as documented in this study. It is known that the EGP and the Rayner complex shared metamorphic history (Fig. 10.7) (Harley, 2003; Morrissey et al., 2015, 2016; Dasgupta et al, 2013, 2017) and these magmatic rocks put additional constraints on the proposed transcontinental correlation. Despite the similarities, the UHT metamorphism at ca. 1000 Ma is not reported from the Rayner Province and thus any correlation that exists must be initiated after ca. 970 Ma (the emplacement age of charnockite magma in the present study). The pre-970 Ma evolutionary history of the EGP appeared to be unrelated to that of the Rayner Province (Bose and Dasgupta, 2018).

The UHT metamorphosed and subsequently cooled crust of the Visakhapatnam domain experienced a granulite-grade reworking during ca. 950–900 Ma (Das et al., 2011; Bose et al., 2011). This reworking has been petrologically constrained by the breakdown of cordierite into a symplectite of orthopyroxene + sillimanite + quartz (Das et al., 2011). Although no clear textural evidence is present, the spread of geochronological data in the range of ca. 960 to 900 Ma recorded both from zircon and monazite grains in the studied rocks provides clues regarding the possibility of granulite-grade reworking in the Phulbani domain. The $^{207}Pb/^{206}Pb$ weighted mean age of ca. 949 Ma obtained from zircon overgrowth and ca. 912−960 Ma spot dates from overgrowth and inwardly projected convolute zones of zircon of the migmatitic felsic gneiss and coarse-grained charnockite might represent later phase of metamorphism within this time span. Despite the constraints of age data obtained from a limited number of samples used in this study, it is evident that the Phulbani and the Visakhapatnam domains evolved as a single entity during the ca. 1000-900 Ma time frame. It contradicts the domain-based classification of the EGP as proposed by Dobmeier and Raith (2003). Major shear zones like the Nagavalli shear zone and the Ranipathar shear zone are thus considered intracratonic resulting thermal activities during ca. 780 Ma and ca. 551-535 Ma and therefore cannot be considered domain boundaries.

The R-EG orogen joined cratonic India and east Antarctica during formation of the Neoproterozoic supercontinent Rodinia through prolonged subduction-accretion (Dasgupta et al., 2013, 2017; Bose and Dasgupta 2018 and references therein). Opinions are divergent regarding the position of India within the framework of Rodinia. While some workers argue that India was indeed a part of Rodinia (Li et al., 2008), others believe that India and a part of east Antarctica

(Napier complex+Rayner complex+Ruker terrane) were never became a part of Rodinia (Cawood and Pisarevsky, 2017; Liu et al., 2013; Merdith et al., 2017; Pisarevsky et al., 2014). Despite this, a shared tectonothermal history between the EGP and the Rayner Complex during ca. 1000-900 time-span is agreed upon in all the models. The tectonometamorphic history from the Phulbani domain also testifies this.