Chapter VII

Petrogenetic and Geodynamic Discussion

7.1 Introduction

Protolith or source and magma play leading role in the evolution of both continental and oceanic crusts, which potentially characterize the magma types, evolution through number of processes and tectonic environments. In other words, protolith is an important component in the generation of magma, which plays a vital role in determining the original chemistry of magma before being modified through later magmatic and tectonic processes acted upon them. These happen after attainment of rheological critical melt percentage (RCMP), the melt gets separated from its source or place of its origin, and ascends and interacts with overlying crustal and/or underlying mafic magmatic components leading to an enigmatic origin of magmatic rocks. However, combination of operative magma chamber processes such as fractional crystallization, assimilation, magma mixing-mingling, liquid immiscibility and many other subsidiary processes may significantly modify the composition of magma.

Most of the Proterozoic mobile belt in the world are either formed by subduction of Archean continental crust or syn-to-post-collisional product of Archean blocks after subduction or could be part of an intracratonic extensional tectonic setting. The common features of such tectonism are production of high-K, I-type metaluminous to peraluminous S-type and post-orogenic to anorogenic A-type granitoids representing bimodal mafic-felsic magmatism.

7.2 Field evidences

The studied intrusive and extrusive rocks of Dudhi gneissic complex are most significant lithounits forming integral part of Mahakoshal Belt in the northern most flank of Central India Tectonic Zone (CITZ). Most of the hidden histories of undeformed felsic plutonic to highly deformed granite gneiss, volcano-sedimentary and post-plutonic mafic dyke rocks association are self-explanatory to their evolution in time and space, which are further corroborated by petrographic, phase-petrological, geochemical and geochronological evidences.
Subtle field observations on the rocks of Sonbhadra district, Uttar Pradesh and also in and around the Sidhi area of Madhya Pradesh which constitute an integral part of Mahakoshal Belt, have been carried out and relationships between the lithounits have been established. Field evidences thus documented have led to revise the stratigraphy of Mahakoshal Belt, which are discussed below in chronological order of geological events.

7.2.1 Basement of Mahakoshal Belt

In the Sidhi area, large to small sized angular xenoliths of pink granite and porphyritic rhyolite (Fig. 7.1a-c) respectively, are found hosted into the massive, fine grained melanocratic volcanics, of Mahakoshal Belt, which were caught up most probably by eruptive volcanics as fragments of Neoarchean basement (Bundelkhand craton). Such xenoliths are not yet reported from any pluton of Mahakoshal Belt, which suggest possible involvement of basement in the generation of later plutonic bodies. However the surface expression of source region in the granites is not a pre-requisite for recognizing the protoliths (e.g. Wall et al., 1987).

Fig. 7.1 (a) Xenoliths of porphyritic rhyolite in metavolcanics of Mahakoshal Group in and around Sidhi area, M.P, (b) Closer view of the porphyritic rhyolite, (c) pink granite of Bundelkhand craton hosted within the same outcrop of metavolcanics.
In the Dudhi gneissic complex, vast exposure of mafic-felsic volcano-plutonic rocks is mainly comprised of migmatized Dudhi granite gneiss (DG) Katoli granitoids (KG), Harnakachar granitoids (HG), Raspahari granitoids (RG) and massive to porphyritic volcanics, occurred in the south of the Renukoot locality of Sonbhadra district. A local fault is passing through these rocks in the north that separates them from northerly exposed lithounits known as Dudhi fault. The northern part of rocks is dominantly made up of sedimentary lithounits in which stock-like felsic bodies Jhirgadandi (JG), Nerueadamer (NG) and Tumiya (TG) plutons have intruded.

The Dudhi granite gneiss (DG) is highly deformed migmatized forming leucosome and melanosome, which are folded and banded. These features have been thought to be mobilized products of basement rocks of Mahakoshal Belt, equivalent to the Chotanagpur gneissic rocks of Jharkhand (Yadav, 1978) and also to granitoid suits of Bundelkhand craton (Ramakrishnan and Vaidyapanadhan, 2008). Conventional wisdom of considering the migmatized-like bodies as basement rocks is not appropriate because of fact that DG has no unconformity relation with any of the exposed lithounits rather it is tectonically juxtaposed with them. Moreover, DG has gradational and diffused contacts with Katoli granitoid pluton. The DG is highly deformed near the Dudhi fault system but away from the fault zone the degree of deformation decreases. It is thus suggested that DG does not represent a basement of Mahakoshal Belt rather has experienced extensive deformational phenomena because of its proximity to fault system.

7.2.2 Evidences of magma intrusion, assimilation, mingling, mixing and undercooling of mafic to hybrid magma globules

Enclaves in granitoids serve as potential tool to understand the magma chamber processes and dynamics (e.g. Kumar, 2010a). French term enclaves cover all kind of lithic materials hosted in granitoids (Didier, 1973, Vernon, 1983). Enclaves could be solid-fragments of country rocks or deeper-derived lithology, known as xenoliths, which signify interaction of solids with granitic liquid, and are genetically unrelated to the host granitoids. Unmelted source lithology and refractory (residual) material left after partial melting can be recognized by peritectic mineral assemblages and partial melting textures. However, in case of xenolith-melt interaction marginal reaction signature may prevail depending upon thermal front of melt. Liquid-liquid or crystal-charged liquids of diverse compositions may interact, mingled and undercooled to form magmatic globules of mafic
to hybrid enclaves, which are an indicative of open magma chamber system intermittently recharged with mafic or hybrid magma into crystallizing felsic magma. Cumulate or autolith may appear as microgranular enclaves but they can be distinguished by typical cumulate textures and grain size differences.

All the granitoids plutons of Dudhi gneissic complex intrude either metasedimentary lithounits or meta-volcanics, except to Dudhi granite gneiss (DG), which has gradational and diffused contacts with Katoli granitoids. The JG pluton intrudes the siliceous cherty-limestone and metapelites as evident by the presence of country-rock xenoliths mostly confined at the margin of the pluton, and core part of the pluton is free from such xenolith. These xenoliths cannot represent the unmelted or residual part of source regions as they lack partial melting texture and, moreover such bigger-sized xenoliths can be transported rheologically from source to sink region.

Fine-grained marginal border facies rocks of JG were not observed. Moreover, there are large mineral grain differences between the JE and host JG, and JE clearly lack cumulate textures. It is thus unlikely that JE represent cumulates or autoliths of host JG, as argued elsewhere (Kumar et al., 2004a). Country-rock xenoliths of metasedimentary origin are confined to marginal parts of the JG pluton which are sometimes ruptured along the foliation planes because of entrained granitic melts at emplacement level. The JE exhibit rounded, elongated and sometimes highly stretched (spindle-shaped) geometry on 2-D outcrops. Interestingly JE of diverse shapes and sizes occur on a single outcrop, which strongly suggests morphometric modification of JE, as a typical feature of magma mingling and flowage. These are largely governed by linear to chaotic processes of mafic and felsic magma interactions in plutonic environment (e.g. Vernon et al., 1988; Wiebe and Collins, 1998; Kumar et al., 2004a). During progressive chaotic mingling and flowage induced by forced convection JE might get mechanically disaggregated and dissolved within the host magma (e.g. De Campos et al., 2011) leading to complete homogenization (Kouchi and Sunagawa, 1983).

Phenocrysts in fine- or medium-grained groundmass or in coarse-grained rocks are stable and crystallized product in host magma whereas xenocrysts are unstable (disequillibrated) phases in a new host (hybrid) magma environment. K-feldspar xenocrysts in hybrid JE are identical, except their partly dissolved crystal outlines, to those of the host granitoids, which are unaffected by deformation and recrystallization
because of relative movement of crystal mush and melt in both ME and host granitoids (e.g. Vernon et al., 1988; Kumar et al., 2004a). Gradational contacts between diorites and granitoids and hybridized features of a quartz diorite, similar to those noted for hybrid JE, oppose their co-magmatic nature and rather favour mixing between coeval mafic and felsic magmas. Presence of mafic-felsic xenocrysts in hybrid diorite and JE is indeed indicative of mechanical transfer of minerals during interaction of partly crystalline mafic-felsic magmas. It resulted in the formation of a (high-T) hybrid magma zone where resorption or partial dissolution of captured phenocrysts occurred leading to formation of anhedral and rounded xenocrysts over which some minerals could have precipitated (Reid et al., 1983; Vernon, 1983; Hibbard, 1995; Kumar et al., 2004a). Near the margin of the pluton JE are stretched and elongated along the contact, which can be attributed to emplacement-related shear flow structure while the JE and magmas had plastic (non-Newtonian) rheology (e.g. Kumar, 2010a and references therein). The JE occurring close to the pluton margin are usually accompanied by leucocratic felsic and pegmatitic veins and dykelets which were probably formed by devolatization and decompression of pluton closer to the contact with country-rocks (e.g. Štěmplok et al., 2008).

The JE exhibit sharp contacts with host granitoids but commonly lack typical chilled margins, which are expected when high-T JE melt comes into contact with relatively cooler host granitoid magma. Sharp contacts of JE indeed suggest undercooling of the mafic magma against the cooler granitoid host (e.g. Vernon et al., 1988; Kumar et al., 2004a). However, quenched texture of JE might have been erased gradually by re-crystallization common in slowly cooled plutons (e.g. Wall et al., 1987). Alternatively, chilled margin could not be developed because of low thermal contrast between mafic and felsic magma mingling system (e.g. Hibbard, 1995) or small size of JE that could be uniformly undercooled developing equigranular fine-grained texture. It is less likely to form small JE with sharp contacts upon disaggregation of bigger JE during progressive mixing because of fact that disaggregation of JE will preferentially produce diffuse boundaries rather than sharp ones. Intrusion of volumetrically subordinate amount of mafic (diorite) to hybrid JE melts into host JG of intermediate composition would lead to mingling, and it would break into pillow-like masses achieving thermal equilibration within hours above the liquidus of the host magma (e.g. Eichelberger et al., 2000). Mafic to hybrid JE occur even at a single outcrop of JG, but order of their injection, interaction or mingling cannot be recognized (e.g. Vernon et al., 1988, Kumar, 2010a). However,
their occurrences in a given JG mass could be controlled by whole-body convection of magma chamber (e.g. Wiebe et al., 1997; Kumar, 2010a).

Southeast of the JG pluton, two-mica leucogranite (NG and TG) plutons intrude the phyllite and slate respectively. The TG pluton forms laccolith structure and consequently the overlying slates have undergone updoming folded and without having any signature of thermal effect on slate (Fig. 7.2a, b). TG is devoid of xenolith but surmicaceous enclaves and angular xenoliths of phyllite are abundant in NG pluton. Surmicaceous enclaves most likely represent residual from source region (e.g Didier, 1973). Andalusite crystals developed in phyllites as a product of low temperature contact metamorphism (Fig. 7.2c, d), as also inferred by Srivastava (1977). Two mica bearing TG pluton is cross-cut by number of tourmaline bearing quartzo-feldsapthic pegmatites (Fig. 2.9e, f), which might have evolved from water-saturated granite magma just on the verge of generating hydrothermal fluids when activity of boron was very high and hence borosilicates (tourmaline) got precipitated.

Abrupt termination of sedimentary lithounits against Dudhi granite gneiss and associated magmatic lithounits in the south of study area (Fig. 2.3), is an indicative of
tectonic contact relationship. In the southern part, vast bimodal mafic-felsic magmatic rocks include HG, DG, KG, RG plutons, massive to porphyritic volcanics, and sedimentary lithounits. HG pluton exhibits diffused contacts with feldspathic meta-quartzite and is devoid of country-rock xenolith (Fig. 2.4d, e). Xenoliths of hornblende (hbl) diorite are however present in HG (Fig. 2.4f), which are unrepresented by country-rock lithology. It is therefore likely that hbl-diorite xenoliths represent deeper-derived lithology. The K-feldspar phenocrysts have precipitated in the host arkosic quartzite because of fine clasts of K-feldspar ingredient and thermal input from intruding HG granite melt. Alternatively, K-diffusion during intrusion of melt has induced the growth of K-feldspar crystals dominantly near the diffusive contact but diminished gradually towards interior of the pluton.

Intrusive nature of KG pluton is evident from xenoliths of schist and metavolcanics country rocks (Fig. 2.6b, c). Xenoliths still preserve original metamorphic fabric i.e. schistocity. Metavolcanic xenoliths armoured with quartz rims (Fig. 2.6c) are suggestive of partial digestion of xenolith during thermal erosion by KG melt. Only leucocratic variety of KG contains xenolith, which indicates xenolith incorporation at apical part of the pluton during late stage of KG evolution. Later the KG magma chamber was cross-cut by numerous mafic dykes. Mesocratic to leucocratic varieties of KG contain mafic to hybrid microgranular enclaves (Fig. 2.6d, e), which represent mafic to hybrid magma globules mingled and undercooled against cooler host granitoids in plutonic setting (e.g. Vernon, 1983; Kumar 2010a).

To the east-southeast of KG pluton, leucocratic variety of RG is emplaced at subvolcanic level as evident from abundant xenoliths of metavolcanics hosted therein (Fig. 2.8a). Swarms of xenoliths are aligned in the direction of magmatic flowage of RG melt (e.g. Phillips, 1968). Gathering and alignment of metavolcanic xenoliths might have occurred at an early stage of RG magma flowage before it got consolidated. These xenoliths resemble with microgranular enclaves as reported elsewhere (Phillips, 1968; Vernon, 1983, 1984). Due to partial assimilation of xenolith by RG melt sigmoidal and lenticular morphology were developed, which are unrelated to post magmatic changes.
Massive to porphyritic volcanic lithounits are intimately associated with alternate sedimentary sequence having diffused and sub-parallel contacts (Fig. 7.3). These volcanics do not contain xenoliths of country rock which suggests quiescence nature of volcanic eruption. However, volcanics have passed through the basement, Neoarchean Bundelkhand granites, and therefore some xenoliths of granite and rhyolite were caught up enroute. The porphyritic volcanics contain poikilitic K-feldspar phenocrysts with biotite inclusion, which are undoubtly of the magmatic origin not the porphyroblasts (Vernon, 1986). These phenocrysts are aligned in layers on the floor of chamber whereas at higher level phenocrysts are randomly oriented which point to lack of thermal convection in the chamber.

### 7.2.3 Magnetite to ilmenite series of granitoids

The studied granitoid plutons in the Dudhi gneissic complex of Mahakoshal Belt have shown affinity with both magnetite (I-type) and ilmenite (S-type) series of granitoids (Ishihara, 1977) corresponding to oxidized and reduced type of granitoids (Fig. 7.4) respectively (Takagi and Tsukimura, 1997). However, all magnetite and ilmenite series of granitoids are not essentially equivalent to metaluminous I- and peraluminous S-type granitoids respectively (Chappell and White, 1974; Takahashi et al., 1980). JG pluton is
highly oxidized as compared to other granitoid plutons of Mahakoshal Belt. JG interacted with cherty sediments and therefore did not reduce to ilmenite series. The core part of the pluton is mildly reduced to ilmenite series because of lower quantity of ferromagnesian minerals as compared to the marginal parts. Granitoids of NG, TG and RG plutons belong to ilmenite series, reduced type of granitoids which are highly peraluminous (S-type) and leucocratic in nature because of their derivation from sedimentary protoliths. About 30% of HG and 50% of KG plutons correspond to magnetite series, oxidized type of granitoids and remaining to ilmenites series, reduced-type granitoids because the magma has interacted with arkosic quartzite and mica-schist country rocks. Since fractionated leucocratic variety of KG is depleted in ferromagnesian minerals, and therefore measures low MS value more akin to ilmenite series granitoids. Tectonism such as regional scale shearing and deformational activities acted upon granitoids can erased out the magnetic properties of granitoids leading to ilmenite series granitoids as displayed by Dudhi granite gneiss (DG).
The volcanic lithounits are suggestive of ilmenite-type of rocks because volcanics assimilated the sedimentary lithounits which might have reduced their magnetic susceptibility (Fig. 7.4).

Based on observed proportions of magnetite to ilmenite series of the granitoids it can be suggested that the felsic magmatism in Mahakoshal Belt was evolving from oxidized (magnetite series) to reduced (ilmenite series) magmatic environment mostly intrinsic to source region except to those local reduction with country rock and tectonic processes (e.g. Singh and Kumar, 2005; Kumar and Singh, 2008). Ishihara et al. (2002) have suggested that muscovite bearing granitoids (NG and TG) showing ilmenite series character may evolve in syn-collisional tectonic setting.

7.2.4 Tectonic scenario and questions remained unanswered

From the field relationships between volcano-sedimentary and felsic plutonism of Dudhi gneissic complex, valid tectonic model cannot be understood because of lack of some observations. The presences of alternate sequence of volcano-sedimentary lithounits may suggest an existence of rift-like tectonic environment. However, absence of carbonitites, lamprophyres and alkaline igneous complexes in the Dudhi gneissic complex does not essentially support the rift like tectonic model. These rocks are however reported from other regions of Mahakoshal Belt which were typically formed in rift-related tectonic setting (e.g Roy and Bandyopadhyay, 1990b). These alkaline igneous suites could have been formed at later stages (~1500 Ma).

- Two-mica bearing NG and TG plutons are products of typical collisional-related tectonic environment, then in which way collisional and extensional-related magmatic rocks are associated to each other?
- Where does subduction-related calc-alkaline granitoids exist?
- Dudhi granite gneiss (DG) is highly deformed and migmatized. Does the DG actually represent basement for Mahakoshal Group of rocks?

7.2.5 Pre to syn sedimentation of magmatism in Mahakoshal Belt

The sedimentary deposits in the Mahakoshal Belt appear to have formed in two-stages. First sedimentation occurred before the intrusion of collisional-related leucogranites (NG and TG). Second sedimentary process operated synchronous with
volcanism in an extensional tectonic regime. However, timing of collision and extension has yet to be established which will throw light on actual relationships between extension and collisional related vast magmatism, which formed the Mahakoshal Belt.

7.3 Petrographic features of genetic value

Mineral assemblages and textural features of magmatic rocks can indicate the nature of protolith, physical conditions of crystallization, and the operating processes during magmatic regime and sub-solidus environment. Modal abundance of constituting minerals of rocks can suggest type of igneous series and likely geochemical features. Despite of crustal and local scale shear deformations, the studied granitoid plutons still preserved magmatic textures except migmatised Dudhi granite gneiss (DG) and massive volcanic lithotypes.

7.3.1 Crystallization of HG magma and its interaction with enclaves

The HG are porphyritic, coarse-grained and contain phenocrysts of K-feldspar. Phenocrysts might have crystallized at depth prior to the emplacement of magma in the chamber, and then slowly cooled in plutonic environment forming coarse-grained matrix. The hbl-pl-Kf-qz assemblage and a few garnets along with few minor phases such as zircon and apatite suggest relatively dry nature of HG melt (Fig. 4.1a, b). Absence of sphene and magnetite in paragenetic sequence might have facilitated crystallization of Ti-amphibole at high temperature in reduced magmatic environment (e.g. Ishihara and Imai, 2014), consistent with dominance of 70% ilmenite series of HG (Fig. 7.4). However, HG are strongly metaluminous, which contradicts to ilmenite series nature of granitoids. Garnet bearing HG correspond to magnetite series which could be due to the occurrence of Fe-rich garnet rather than magnetite. Wavy contact between garnet and plagioclase and absence of corona texture suggest typical magmatic origin. Biotite has precipitated late in sequence along the fracture system of garnet.

Folded biotite (ductile) and distorted twinned lamellae of plagioclase (Fig. 4.2a, b) strongly suggest deformation synchronous with crystallization. This is further supported by discontinuity of such deformation in the surrounding minerals. Some growth and deformed twins in plagioclase suggest abrupt termination of multiple twinned plagioclase and lenticular type of abrupt termination formed during deformational activity.
respectively (Vernon, 2004). Both the features may or may not occur within a single crystal of plagioclase depending upon the semi-solid or solid-state deformation.

HG host two types of enclave; one is garnet bearing microgranular enclave and another is garnet-free hornblende diorite xenolith (Fig. 4.2e, f). The garnet in enclave (HE) is magmatic in origin, which exhibits crude alignment with biotite flakes showing magmatic flowage. On the other hand, the xenolith of hornblende diorite is fine grained which has crystallized at early stage.

Garnet bearing HE does not exhibit any of the restitic, cumulate, mafic-to-hybrid microgranular enclave types. However, it is also doubtful to be called xenolithic since it has diffused or corroded contacts with the host HG. Although, HE has similar mineral assemblages as noted for host HG but has more modal proportion of mafic minerals. The presence of garnet in the host HG is relatively lesser than the HE that may have been mechanically transferred from HE during interaction between them and new layer of garnet over incorporated garnet in a new felsic magmatic system of HG might have taken place.

7.3.2 Evidence of JE mingling and syn-crystallization in JG

Microgranular enclave hosted in JG clearly lacks textural features of restite, cumulate or xenolithic origin. Rather they exhibit fine-to medium-grained equigranular to hybridized (xenocrystic) signatures having mineral assemblage identical to that observed in host granitoid, which could mislead to cogenetic (cognate) relationships as early-formed fine- to medium-grained border facies of the JG pluton itself. Distinctly higher modal contents of ferromagnesian minerals in mafic JE compared to host JG and variable amount of felsic-mafic xenocrysts in hybrid JE indeed suggest interaction of crystal-charged mafic and felsic magmas (Bora, 2013). These xenocrysts formed during progressive mingling and mixing events, which may indicate temperature of homogenization (e.g. Van Der Laan and Wyllie, 1993). Therefore mafic xenocrysts destabilized within JE magma cannot be considered as disrupted restite (e.g. Reid and Hamilton, 1987; Kumar and Rino, 2006). Presence of pyroxene as resorbed or independent grains in JE strongly points to dry and mafic nature of the JE magma (similar to diorite) prior to its interaction with the coeval host granitic melts under plutonic conditions. Mingling of hydrous and relatively cooler JG magmas with dry and hotter JE melts could have produced varying amounts of vapour phase which induced the
precipitation of hydrous ferromagnesian phases (amphibole and biotite) in JE at the expense of pyroxenes (e.g. Kumar, 2010). Absence of needle-shaped apatite in JE appears to be a combined effect of low thermal contrast (100–150 °C) between JE and JG melts, phosphorous-depleted JE melts and resident time of JE liquid essentially controlled by the size of the JE (e.g. Kumar and Rino, 2006 and references therein).

7.3.3 Crystallization of KG magma and its interaction with enclaves

Mafic to felsic mineral proportions are changing in the rock sequence of KG pluton (Fig. 4.4a-f), which indicate evolution of KG through fractional differentiation. Diorite comprises of cpx-hbl-pl and is devoid of mag-ap-ttn-zrn, which together indicate dry and reduced (low $f_{O_2}$) condition of melt at an early stage of dioritic magma evolution. With the progress of crystallization, $H_2O$ content of magma and $f_{O_2}$ increased in KG melt suitable to crystallize hydrous ferromagnesian minerals (hbl-bt), sphene and magnetite. The crystallization of euhedral sphene and Fe-Ti oxides consumed the available oxygen in the melt leading to again reduced condition of KG melt. Therefore hydrous ferromagnesian minerals (hbl-bt) are found abundantly in mesocratic, melanocratic varieties of KG and then progressively declined in leucocratic variety of KG, during fractional differentiation. With the progress of crystallization, amount of K-feldspar and quartz has increased. Zircon content has lowered in leucocratic variety of KG because Zr-solubility in felsic melt is sensitive to peraluminosity of granitic melt (Watson, 1974). Changing proportion of mafic to felsic minerals and Fe-Ti oxide phases has therefore greatly influenced the proportion of magnetite to ilmenite series granitoids of KG. This resulted in the formation of almost 50% ilmenite series (reduced-type) granitoids of KG pluton (Fig. 7.4).

Meyrmekitic, vermicular, perthitic textures are more common in the late evolutionary stages of KG melt (Fig. 4.4e, f), which were formed under sub-solidus conditions. In addition some perthitic K-feldspars showing Baveno twining have experienced syncrystallization ductile deformation when the rock was sufficiently heated below the solidus. It is also possible that regional-scale shearing deformational activities were synchronous to granitoid pluton emplacement, which is corroborated by the presence of plagioclase with twinned lamellae having finer tips during syncrystallization tectonic activity.
7.3.4 Migmatized nature of DG granite gneiss

Dudhi granite gneiss (DG) displays migmatite textures at outcrop and microscopic levels. There are evidences of melt segregation but melt has not moved out of migmatized system, and therefore can be termed as insitu leucosomes that have also minor amount of mafic selvedges (melanosomes), as described elsewhere (Sawyer, 1996). Leucosomes are defined by quartzofeldspathic minerals whereas melanosomes are mafic (biotites) rich with occasional amphiboles. Some of the plagioclase in the leucosome could be residuum of DG protolith. It is inferred that DG has suffered low grade of metamorphism: a condition prevailed in the beginning of melting, which formed the patches of migmatites.

7.3.5 Magmatic processes, and protolith assessment of RG pluton

The RG pluton is mainly comprised of pl-kf-bt-qz±ms±gt-zrn-ilm±ap assemblages, which were typically formed from crustal-derived magma. The modal amount of plagioclase and biotite decreases with increasing content of quartz and K-feldspar (Fig. 4.7a-d), which indicate evolution of granitoids by fractional differentiation of a parental RG melt. The presence of garnet having K-feldspar inclusion (Fig. 4.8a) suggests direct involvement of sedimentary protolith in the production of RG melt, which contains Al-rich restitic component but eliminated from fractionated felsic melt (e.g Chappell et al., 1987). In the melting event of sedimentary protolith (Deer et al., 1997), the RG melt forming reaction can be written as:

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\text{Biotite + plagioclase + aluminosilicates + quartz = garnet + K-feldspar + melt}
\]

Absence of magnetite and presence of ilmenite suggest reduced nature of RG melt which is intrinsic to sedimentary source region and hence corresponds to ilmenite series granitoids (e.g. Ishihara, 1977; Kumar and Singh, 2008) consistent with observed MS value. Ilmenite series RG typically represent S-type granitoids (Whalen and Chappell, 1988). Absence of titanite indicates low temperature and low \( f_{O_2} \) (reduced) conditions for RG melt. It is likely that the observed detrital zircons enclosed within biotite (Fig. 7.5) might have inherited from sedimentary protolith involved in the generation of RG melt.

Low modal amount of apatite in RG suggests that the source was depleted in Ca component. It is generally observed that solubility of P in peraluminous granites is higher and progressively increased during fractionation which may or may not crystallize apatite depending upon early monazite crystallization (Broska and Petřík, 2014).
Monazite is not crystallized in RG melt, and therefore rule out former situation. On the other hand if the sedimentary protolith could be phosphorous bearing carbonaceous metasediments then the RG melt would have enriched in P under prevailing condition, however, apatite is rare or absent in RG melt. Thus, it is more likely that a metapelitic sedimentary protolith might have been involved in the generation of RG melt which is consistency with the equation of melting given above.

Hollocrystalline nature of microgranular enclave (RE) hosted in RG is comprised dominantly of plagioclase, biotite, K-feldspar and quartz, with abundant acicular apatite and stubby magnetite (Fig. 4.8b), which suggest hydrous nature of enclave magma (relatively mafic) which undercooled as RE globules against relatively cooler RG host magma (e.g. Kumar, 2010a and references therein). The presence of magnetite in RE suggests prevailing oxidizing condition at the time of magma interaction with felsic RG host, which is a typical feature of open magma system (Kumar, 2010a). On the other hand, garnet bearing xenolith hosted in RG might have thermally metamorphosed and therefore have formed garnet (e.g. Mass et al., 2001). It is likely that xenolith may represent either restite (e.g White et al., 1999) or deeper-derived exhumed lithology (e.g. Mass et al., 2001).
7.3.6 Magmatic processes and protolith assessment of NG and TG melts

The presence of two-mica (bt-ms) in NG and TG plutons indicates formation of \( H_2O \)-saturated leucocratic melt in collisional environment (Wyllie, 1977; Barbarin, 1990) which was derived from metasedimentary (metapelitic or metagrewacke) sources. The generation of lecogranite melt is common feature of collisional tectonics, but the question arises when did it happen and what is its relationships with other mafic-felsic magmatic units of Mahakoshal Belt? These questions will be answered satisfactory while discussing geochronological data.

Hydrothermal alteration of NG and TG is indicated by sericitization of plagioclase. The presence of small quartz “blebs” inclusion within K-feldspar and plagioclase (Fig. 4.9a-d) strongly suggests greater undercooling which promoted higher nucleation rate of quartz (e.g. Hibbard, 1995). Alternatively cotectic crystallization of feldspars and quartz in silica oversaturated melt leads to formation of rounded to subrounded quartz blebs, which were later grown to vermicular and tiny quartz inclusions along the K-feldspar boundaries close to the waning stage of crystallization at ternary minima of Qz-Or-Ab system (Vogt, 1921; Dunham, 1965; Barker, 1970; Hughes, 1971). Thus, the silica saturated felsic magmatic system crystallizes quartz and feldspars cotectically up to the intergrowth stage (Hibbard, 1995). Poikilitic and perthitic textures are common in K-feldspar, which primarily depend upon rate of diffusion of elements at crystal growth site and at subsolvus condition respectively. The latter condition is achieved either through addition of Ca component in the solid-solution system of albitic plagioclase or due to increased \( fH_2O \) which lowers the temperature of crystallizing K-feldspar (Morse, 1970 in Philpotts, 1994). Since both NG and TG are poor in Ca component, and therefore the latter condition is more likely than the former, equivocally evident by the presence of two-mica (bt-ms) in these granitoid plutons.

7.3.7 Textural similarities among the granitoid plutons

Poikilitic, strings and flames of perthites, cross-hatched, myrmekitic and vermicular textures along with lamellar, albitic, Carlsbad, pericline, simple twinning in feldspars are common features observed in granitoid plutons of Dudhi gneissic complex. The perthitic textures are developed under subsolvus condition whereas myrmekites represent intergrowth texture developed along grain boundaries of K-feldspar and quartz, and the absence of any flowage or foliation pattern indicative of complete crystallization of felsic
melt at eutectic (Wang, et al., 2001). They can also be formed during syndeformational regimes in which microaplitic vein may show mantle-like crystallization parallel to foliation suggest myrmekitic exsolution perpendicular to the maximum shortening (e.g. Hibbard, 1979, 1987; Hutton, 1988) along the grain boundary during deformational regime as observed in DG pluton. Such textural features are also common in HG and KG plutons, which suggest ductile deformation of plutons probably occurred during shearing effect of CITZ, as inferred from folded biotite crystals and distorted multiple twinned lamellae of plagioclase in HG (Fig. 4.2a-b), and wavy Baveno twinned plane in K-feldspar of KG (Fig. 4.4e).

7.3.8 Volcanism and assimilation with country rocks

The volcanic lithounits are massive to porphyritic in nature. Massive volcanic lithounits are foliated and sheared composed of amphibole (relict pyroxene; Fig. 4.10c), biotite, plagioclase, K-feldspar and quartz, which indicate anhydrous to hydrous nature of mafic melt contaminated with crustal components during ascent and eruption. Porphyritic volcanics having phenocrysts of pokilitic K-feldspars and randomly oriented unzoned plagioclases (Fig. 4.10d, e) suggest their typical magmatic origin not the porphyroblasts (Vernon, 1986). These K-feldspars are partially rimmed by precipitation of new K-feldspar growth, most probably towards the end of magmatic regime forming rapakivi like (sensu lato) textures during assimilation and thermal contact (e.g. Hibbard, 1995) with country rock (quartzite) which had undergone crystallization against crystal edges of the K-feldspar (Fig. 4.10d). It contains inclusion of plagioclase which may suggest turbulent flow experienced by plagioclase during growth of K-feldspar (Hibbard, 1965). Growth, nucleation, and diffusion rate control the formation of poikilitic inclusions in phenocrysts (Vernon, 2004). Smaller inclusions than the surrounding matrix suggest early crystallization of inclusion, whereas matrix crystals continued to grow at later stages (Higgins and Kawchi, 1977). This does not hold truth as inclusions in plagioclase are relatively bigger than the fine grained matrix. Based on textural features, at least two-stage of crystallization of volcanics can be suggested. This inference is clearly indicated by randomly oriented plagioclase phenocrysts embedded in the fine grained groundmass composed of biotite, K-feldspar and quartz (Fig. 4.10c). These plagioclases have irregular corroded boundaries (not sharp with matrix) and contain more prominent inclusions and interfingering relation with groundmass biotites (Fig. 4.10e) as similarly observed elsewhere (e.g. Pitcher and Berger, 1972) which suggest that the phenocryst continued to
grow and thus incorporated the surrounding biotites flakes in its marginal boundaries (Vernon, 1986).

The volcanics of Dudhi gneissic complex are intimately associated with quartzite, which has angular fragments of quartz and K-feldspar embedded into mature matrix of quartz (Fig. 4.10f). It suggests the existence of energy condition for deposition of sediments, which were derived from proximal granitic provenance. These sediments were almost penecontemporaneous with volcanism, and have reacted with them to form diffused contacts.

7.3.9 Porphyritic rhyolitic xenoliths in mafic volcanics of Sidhi area: evidence of fragments of Bundelkhand craton basement

![Fig. 7.6 Porphyritic rhyolite xenolith caught up in metavolcanics that is composed of euhedral amphibole phenocrysts embedded in fine grained quartzofeldspathic groundmass.](image)

Idiomorphic to subidiomorphic amphibole phenocrysts are embedded in cryptocrystalline felsic (quartz and feldspar) groundmass (Fig. 7.6) forming the porphyritic rhyolite, which is hosted as xenoliths in basalt porphyry (~70% phenocryst of pyroxene embedded in basaltic groundmass). Pyroxene phenocrysts are altered to amphiboles but the relict pyroxene cores are still preserved giving rise to resorbed texture. Amphiboles
are partly chloritized. Some of the plagioclase are altered to sericite. Basaltic groundmass is composed of laths of plagioclase and granular pyroxenes, respectively altered to sericite and amphiboles. Rhyolitic xenolith has shown reaction margins of recrystallized quartz with basalt, which indicates solid-melt interaction forming siliceous margins. Rhyolitic xenolith has resemblance with the rhyolite of Bundelkhand craton.

7.4 Mineral chemical evolution of magmas

7.4.1 Nature of JG melt: evidence from pyroxene composition

The JG pyroxenes are resorbed by amphiboles as evident from conversion of diopside to augite leaching out wollastonite component from primary pyroxene (Fig. 4.12). This has happened because of increasing water content of JG melt facilitating accommodation OH in the structure crystallizing amphiboles at the expense of pyroxene (Bowen, 1928). Thus the clinopyroxene fractionation in JG melt was terminated with the crystallization of amphiboles which replaced the pyroxene on liquidus: a feature similarly observed elsewhere (Dorais, 1990; Kumar, 1996). JG clinopyroxenes are Ca-rich augite and lie on alkaline melt trend (Fig. 4.13) but are low in acmite (Na₄pfu) content suggesting tholeiitic to mildly alkaline nature of parental melt (e.g. Kumar, 1996 and reference therein) exclusively formed in extensional-related tectonic environment.

7.4.2 Redox-state of evolving melts under variable P-T conditions: imprinted on amphibole composition

Amphiboles from JG, DG, HG, enclaves (JE, KE, HE) and dyke (HD) represent calcic-amphiboles of magmatic origin (Fig. 4.16-17a, b). The JG and JE amphiboles have relatively lower Al⁴⁺ content as compared to those of other plutons, which indicate solidification at low-P (ca 2 kbar) shallow crustal level (table 4.2-4.4; Fig. 4.20a). However, amphiboles in other plutons might have crystallized at deeper crustal level (ca 7 kbar; table 4.5-4.6; Fig. 4.20a), which bear relatively high Al⁴⁺ content. In paragenetic sequence crystallization of anorthitic plagioclase prior to amphibole may possibly deplete the Al content in the melt (Fig. 4.18a, b). However, Si₄pfu in amphiboles indeed determines the accommodation of Al at T-site. It has already been stated that amphiboles in JG crystallized as the aH₂O in the melt increased. Amphiboles in other plutons have crystallized first in paragenetic sequence at elevated aH₂O in the melts having relatively lower aSiO₂ because of intermediate composition of melts. Pressure is another factor which
controls the high Al content in T-sites of amphibole structure. *Edinitic-type* of substitution is less common because amphiboles are crystallized prior to plagioclase (e.g. Castro, 1992). However, amphiboles have crystallized cotectically with plagioclase. The varying degree of *tschermakite-type* substitution has been observed in almost all studied amphiboles (Fig. 4.18a). *Tschermakitic-type* of substitution is one of the factors which is responsible for redox-state of the magmas (Phillips et al., 1988), as coupled-substitution particularly at C-site (Phillips et al., 1989).

The redox condition of melts in which amphibole crystallized depends upon dissociation of hydroxyl ion, which causes surplus ferric-state of iron in amphibole, leading to oxidizing environment. It may also be reversible through change of ferric to ferrous by consumption and re-introduction of hydrogen in the melt (Barnes, 1930) in the form of OH⁻ ions in amphiboles that may have to be charge balanced by incorporation of more ions at C-site. This mechanism therefore has led to the formation of JG amphiboles in oxidizing environment as they have relatively higher Fe_total as compared to the amphiboles of other plutons crystallized in relatively lower oxidizing environments. This inference is consistent with the observed magnetic susceptibility (MS) values of JG, which suggested magnetite series (oxidized type) of granitoids. Other plutons have however determined the relatively lower MS values formed in moderately oxidized to reduced type granitoid magma environments. JG amphiboles contain relatively lower Ti-content as compared to those of other plutons (Fig. 4.21), which point to crystallization of JG amphiboles at lower-T than the other amphiboles (e.g. Deer et al., 2007). Experimental results on amphiboles (Blundy and Holland, 1990) further strengthened the inferences of low temperature (JG and JE) and high temperature (HG, DG, KE, HE and HD) amphibole crystallization in their respective melts (Fig. 4.20b). The high Ti-in-amphiboles (HG, DG, KE, HE and HD) possibly suggest early crystallization of amphibole than sphene whereas low Ti-in-amphiboles (JG and JE) is caused by amphibole crystallization after sphene in the evolving JG melt. Since, titanite is crystallized under increased fO₂ condition of magma (Wones 1989), thus high Ti-in-amphiboles (HG, HE, KE, DG and HD) must have formed under low oxidizing condition at high-T followed by crystallization of sphene with progressively increasing fO₂ while cooling of respective host magmas. High Ti-in-amphibole is proportional to K₂O content of host magmas which are found consistent with observed Ti-in-amphiboles and K₂O-content of plutons in the present investigation (Figs. 4.18c, 4.21a). The K₂O content also
depends upon low oxidizing condition of melt which can accommodate Fe$^{2+}$ in amphibole (Dawson and Smith, 1973). Hence, high $f$O$_2$ condition might be a valid reason for crystallization of JG amphiboles with lower Fe$^3$/Fe$^{3+}$+Mg) ratios i.e. beyond the limit of amphiboles (0.40-0.65; table 4.2-4.4) for the use of Al-in-amphiboles barometers (Andreson and Smith, 1995). Fet/(Fet+Mg) ratios of amphiboles in other plutons are in the range of 0.40-0.65 (Table 4.5-4.6), which suggest moderate $f$O$_2$ condition in the evolving mafic-felsic melts (HG, HE, HD, KE, DG). The Ti-in-amphiboles are controlled by couple substitution:

$$Fe^{2+} + OH^- = Fe^{3+} + O_2^- \text{ and } [6]Al + OH^- = Ti^{4+} + O_2^-$$

as suggested by Popp et al. (1990) and Popp and Bryndzia (1992) for amphiboles $>$0.7 Fe$^{3+}$ apfu. In the studied plutons low and high-Ti amphiboles are found below the suggested range, which should be controlled by early or late crystallization of sphene in the oxidation-dehydrogenation of a tschermakite-type of amphibole (e.g. Philliphs et al., 1989). It is therefore inferred that changing redox state of evolving magmas were dominantly controlled by order of crystallization of mineral phases in equilibrium with melt.

7.4.3 Subsolvus granites: evidence from feldspar thermometry

Subsolvus granites are either products of late stage crystallization of magma (e.g. Kovalenko and Kuz'min, 1969; Kovalenko and Kovalenko, 1984; Aubert, 1969; MacKenzie et al., 1988) or could be caused by post-magmatic hydrothermal fluids (e.g. Taylor and Fryer, 1983; Higgins et al., 1985; Nurmi and Haapala, 1986; Que and Allen, 1996; Perez and Boles, 2005; Plümper and Putnis, 2009) as both the processes may occur under subsolvus condition. These two processes responsible for subsolvus growth can be distinguished in petrographical observations. Three major stages are involved for the development of subsolvus granites of hydrothermal origin (Plümper and Putnis, 2009); (i) K-feldspars may get converted into oligoclase and relics of microcline can be observed within secondary oligoclase, (ii) the secondary oligoclase may alter into sericite due to sericitization, and (iii) further K-feldspathization along sericite may take place and hematite may precipitate during hydrothermal activity, and consequently K-feldspar is stained with red colour because of prevailing oxidizing condition. However, such features are absent (except few sericite inclusions hosted in plagioclase of RG pluton), and therefore fresh part of plagioclase was selected for electron microprobe analysis.
The subsolvus granites of studied plutons are of magmatic origin, and have equilibrated in the temperature range of 460 °C-550 °C (table 4.25) that may form intermediate sanidine (e.g. Smith, 1974). Experimental works on Qz-Ab-Or (Tuttle and Bowen, 1958; Pichavant and Manning, 1984) constrained that subsolvus granites are generally formed at non-minima point which are outcome of increasing $f_{H_2O}$ condition of melts (MacKenzie, 1986). All the granitoid plutons (except HG pluton) of Dudhi gneissic complex of Mahakoshal Belt have attained subsolvus conditions marked by perthitic, myrmekitic, vermicular intergrowth textures characteristics of respective K-feldspars in various granitoids. The obtained results of two-feldspar thermometry independent of pressure suggest that the feldspars in plutons continued to cool below subsolvus because of changing partial pressure of water ($P_{H_2O}$) in the evolving magmas. Experimental works carried out elsewhere have suggested that the felsic magma may crystallize in a large temperature range up to extremely below the solidus even below 400°C (e.g. Moody et al., 1985), and it is therefore inferred that these granitoid magmas were cooled under magmatic to submagmatic temperatures and are unaffected by hydrothermal regimes.

Single feldspar thermometry (table 4.27a, b) suggests that HG plagioclase crystallized at higher temperatures (1266-1226°C) as compared to plagioclases of other granitoid plutons. This suggests higher liquidus temperature of HG magma. Plagioclase of JG diorite exhibits temperature (904-884°C) of crystallization closely followed by plagioclase temperatures in monzogranite at rim (855-719°C) to core (648-594°C) of JG pluton. These suggest plagioclase crystallization at higher temperatures that continue to cool even at subsolidus, from margin to core part of the pluton. Fine grained enclaves solidified at higher temperature (undercooled; 1082-826 °C) than slowly cooled coarse grained enclaves hosted in JG pluton. Both hybrid diorite (849-760 °C) and hybrid coarse grained enclaves (818-684°C) gave more-or-less same range of plagioclase crystallization temperatures. Condition of the plagioclase crystallization in both environments is the same. This is probably because of fact that the hybrid (JE) enclave is disaggregated products of hybridized dioritic melt. However, slight temperature difference suggests that plagioclase in the hybrid enclave globules lost temperature quickly while mixing-mingling processes. It is interesting to observed that plagioclase of KE (microgranular enclave; 1209-1111°C) hosted in leucocratic variety of KG (Fig. 2.6e) yielded higher temperature of crystallization than those host KG (837-87°C). The temperature difference
is large between KE and KG plagioclase because mixing was inhibited due to high viscosity caused by high amount of crystal load in the melt. Under such rheological condition quenching and undercooling of KE magma against cooler host magma was not plausible. KE was already crystal-charged (semi-solid) coarse grained before it came into contact with host KG magma, and therefore could not develop acicular apatite crystals. This is further supported by the presence of cylindrical shaped apatite in KE (Fig. 4.5d). Plagioclases from both deformed (1129-1092°C) and undeformed (1083-913°C) DG were crystallized at the same temperatures which indicate that the degree of deformation has not affected the primary temperature of plagioclase crystallization. Plagioclases in highly peraluminous RG (791-761°C), NG (776-643°C) and TG (719-593°C) magmas were crystallized near-solidus (table 27b).

7.4.4 Redox conditions of granitoid magmas: evidence from biotite chemistry

The presence of magnetite and titanite (sphene) co-existing with K-feldspar and biotite may indicate the oxidation conditions of magmatic and late- to post-magmatic stages (Wones and Eugster, 1965, Wones, 1972, 1989, López-Moro and López-Plaza, 2004). The composition (Fe3+ as oxyannite and Fe3+/Fe3+Mg) of biotites can provide relative oxidation state of magma under which they have crystallized. An isobaric (P=2070 bars) experimental calibration of biotite equilibria carried out at various buffers (FMQ, NNO, HM) of granitoid melts (Wones and Eugester, 1965, Wones, 1972) can be used to estimate minimum temperature (T °C) and fugacity of oxygen (fO2) of biotite stability into granitoid melts.

Granitoids of JG, KG, DG and HG plutons contain assemblage (bt-Kf-qtz-mag) required for estimation of fO2 and T of biotite stability in respective granitoid melts. Biotites from granitoids were analysed using electron microprobe, and hence total iron FeO was obtained. Partitioning of Fe3+ and Fe2+ from total iron (FeO) was empirically estimated following the charge-balance method as outlined by Dymek (1983). Biotite from JG, KG and DG appear to have evolved from FMQ to NNO buffers gradually changing oxyannite component of biotites (Fig. 7.7). However, biotites from Jhirgadandi (JE) and Katoli (KE) enclaves are confined to FMQ buffer. This is probably because of fact that enclave globules were undercooled within host granitoid melts and they might have experienced relatively low oxidizing condition of crystallization. Biotites from enclave (HE) hosted in HG have experienced evolution from FMQ to NNO buffers that
are possible when enclave globules have relatively longer resident time of liquid + crystal conditions within the partly crystalline host granitoid melt. This is further collaborated by the observed elevated oxidizing (NNO) condition of biotite stability in HG magma. Biotite composition in enclave may have re-equilibrated (modified) with those of granitoids during prolonged interactions between coeval mafic-felsic magmas (e.g. Kumar and Rino, 2006), and hence biotite composition of enclave may not be appropriate for estimation of T (ºC) and fO2 of enclave magma. Nonetheless, it can be qualitatively inferred that mafic-felsic magma mingling and mixing has occurred in typical open magma system at relatively elevated oxidizing conditions.

Biotite compositions (100xFe/Fe+Mg) from JG, KG, DG and HG have been projected onto isobaric T-fO2 section of experimental biotite equilibria (Wones and Eugster, 1965), which shows a range of fO2 and TºC for these granitoid melts as summarized in the table 7.1. The JG melt exhibits higher oxidizing trend of biotite evolution (FMQ-NNO; log fO2 = 10^{12.41}-10^{12.51} bars; T=880-900ºC) as compared to biotite crystallized in KG melt at relatively lower oxidizing condition (FMQ-NNO; log fO2 = 10^{13.76}-10^{14.16} bars; T=730-820ºC). Biotites in DG melt (FMQ-NNO; log fO2 =
Table 7.1: Summarized geochemical features and chemical parameters of genetic significance of biotites from various granitoids of Dudhi gneissic complex of Mahakoshal Belt.

<table>
<thead>
<tr>
<th>Mineral Assemblage</th>
<th>Jhingra-dandi granitoid</th>
<th>Jhingradandi Enclave</th>
<th>Katoli granitoid</th>
<th>Katoli Enclave</th>
<th>Dudhi granitoid</th>
<th>Rasphahri granitoid</th>
<th>Nerueadamar granitoid</th>
<th>Tumia granitoid</th>
<th>Harunkachar granitoid</th>
<th>Harunkachar Enclave</th>
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<tbody>
<tr>
<td>SiO₂</td>
<td>36.00-38.48</td>
<td>37.42-37.83</td>
<td>34.13-35.47</td>
<td>34.91-36.10</td>
<td>34.66-36.38</td>
<td>34.12-35.37</td>
<td>31.76-34.52</td>
<td>33.89-34.73</td>
<td>34.09-35.43</td>
<td>34.65-35.75</td>
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<td>1.17-1.20</td>
<td>1.47-1.53</td>
<td>1.45-1.54</td>
<td>1.41-1.50</td>
<td>1.50-1.59</td>
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<td>1.63-1.73</td>
<td>1.31-1.38</td>
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<tr>
<td>X₃p (Fe²⁺ Fe³⁺ Mg)</td>
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<td>0.58-0.62</td>
<td>0.52-0.57</td>
<td>0.55-0.59</td>
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<td>TiO₂ (wt %)</td>
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<td>2.70-3.43</td>
<td>2.43-2.63</td>
<td>1.98-3.38</td>
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<td>Moderat</td>
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</tbody>
</table>

A/Al₂O₃: Al₂O₃/CaO+Na₂O+K₂O of biotite; A/Al₂O₃: Al₂O₃/CaO+Na₂O+K₂O of whole rock

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The obtained T and fO₂ can be used to estimate fH₂O of granitoid melts at different buffers (FMQ-NNO) using the equilibrium reaction relation of annite, sanidine and magnetite (Wones, 1972; Wones, 1981). In case of calc-alkaline, metaluminous (I-type) JG, KG, DG and HG melts, activity of magnetite (aFe₂O₃=1.0) and activity of sanidine (a₃₅=0.8) in K-feldspar can be assumed as suggested elsewhere (e.g. Speer, 1980).
Mole fraction of annite ($X_{\text{annite}}$) in octahedral site of biotite (Wones, 1972) of respective granitoids can be used for calculation of $fH_2O$ at different buffers. Intersection of biotite equilibrium curve at FMQ and NNO with the $H_2O$-saturated granodiorite solidus (after Piwinski and Wyllie, 1968) may provide minimum temperature ($T^{ºC}$) and maximum $fH_2O$ (kbar) of granitoid melt. The obtained results of $fH_2O$ (kbar) at different buffers (FMQ-NNO) for JG, KG, DG and HG melts are summarized in the table 7.1.

The estimated $fH_2O$ (0.30 kbar) at 880 ºC for JG melt is substantially lower than those calculated for KG ($fH_2O=1.10$, NNO, $T=730^{ºC}$), DG ($fH_2O=1.00$, NNO, $T=750$ ºC) and HG ($fH_2O=1.00$, NNO, $T=740^{ºC}$) magmas. This is probably because of relatively dry ($H_2O$-undersaturated) conditions of JG magma, which is relatively more metaluminous as compared to KG, DG and HG magmas. However, the estimated solidus temperature ($880^{ºC}$) for JG melt appears to be higher than the real (~800ºC) one. The increased $fH_2O$ (~1.00 kbar) near or at the solidus as noted for KG, DG and HG melts might have built-up high $P_{H_2O}$ because of early and much crystallization of anhydrous rock-forming minerals in respective granitoid melts, which were most likely responsible for the generation of abundant hydrothermal system developing pegmatite and micro-pegmatites as observed in the regions.

7.4.6 Tectonic implications: based on biotite chemistry

Biotite compositions of granitoids can be applied to predict the tectonic setting of host magmas (Abdel-Rahman, 1994) provided the inferred tectonic settings are consistent with field relation, petrologic and geochemical features (Pitcher, 1983, Barbarin, 1999). Compositions of biotite from JG suggest nature of host magma as metaluminous (I-type) formed in subduction setting whereas biotites from NG and TG relate to peraluminous (S-type) magma originated in syn-collisional tectonic setting (Fig. 4.25a-c, 4.26). These inferences are consistent with observed hbl-bt-pl-Kf-qtz-ttn and bt-ms-Kf-pl-qtz-mz assemblage in JG and TG respectively. Compositions of biotites from KG, DG and HG indicate transitional (H-type) nature of host magma as they plot on the boundary separating subduction-related calc-alkaline metaluminous and syn-collisional peraluminous host magmas, which pose problem for correct usage of tectonic discrimination solely based on biotite composition (e.g. Machev et al., 2004). The tectonic issue of granitoid plutons are discussed further in sections 7.5.8
7.4.7 Muscovite composition as an indicator of barren and mineralized granites

Two-mica (ms-bt) granitoids and pegmatites are important sources for the deposition of economic minerals of Sn, Nb, Ta, Li, Be and Rb (e.g. Taylor, 1979; London, 1987; MacKenzie et al., 1988), which typically belong to peraluminous (S-type; Chappel and White, 1974), ilmenite series granitoids (Ishihara, 1977) commonly formed in syn-collisional tectonic setting (Barbarin, 1999). Two-mica NG and TG plutons contain primary muscovites, which belong to primary zinwaldite (Fig. 4.27-4.28). TG muscovite is pure muscovite that is capable to form tin deposits whereas NG muscovite is Li-Fe muscovite and therefore related to barren deposits (Tischendorf et al., 2007). Both NG and TG plutons bear the same mineral assemblages and textural characters but one is barren and another has potential to form Sn deposits respectively. There may be two reasons for barren and mineralizing ability of NG and TG pluons. Firstly, Sn concentration might have increased with increasing degree of differentiation for TG pluton (e.g. MacKenzie et al., 1988) or differential degrees of melting of protolith could be responsible for mineralization. Secondly lack of Sn in NG, may have either increased fO₂ or restite unmixing forming residual mineralogy (surmicaceous enclave common in NG whereas rare or absent in TG) due to low degree of melting or biotite fractionation during evolution of melt (e.g. Stone, 1982). The second possibility is more likely that can explain the lack of Sn in NG due to presence of restite (surmicaceous enclaves) distinctly observed in NG. On the other hand, TG pluton potential for Sn-deposit (not yet reported in the area, and need to be explored) does not contain any such restitic materials rather cross-cut by tourmaline bearing pegmatites that is absent in NG pluton. It is thus inferred that increased fH₂O without entrainment of restitic materials with increasing degrees of melting are responsible for mineralization. This has been concluded elsewhere that the fO₂ in a granitic magma may be more important than the nature of granitic protolith in controlling type of associated mineral deposits (Whalen and Chappell, 1988). This aspect is further discussed while dealing with whole rock geochemistry.

Tin deposit either may be result of partial melting of continental crust in subduction or intracratonic-rift environment (Taylor, 1979). Ilmenite series (S-type) granites in Japan (Ishihara, 1981) have shown signature of cassiterite deposits. Muscovite, occurrence in ilmenite series is one of the essential minerals for acquisition of the high concentration of ore-forming elements (OFEs) by the granitic magma (Eugster, 1985). Experimental work on Sn mineralization has suggested that casseterite deposition depends on the fO₂ and T
of melt. Further, cassiterite can be soluble in supercritical chloride fluids rich system in equilibrium with albite-K-feldspar-muscovite-quartz phases whose temperature ranges between 400 and 600°C ~ at 1.5 kbar in reduced environment (\(SnO_2 \leftrightarrow SnCl^+\), Wilson and Eugster, 1984). The obtained temperature ranges (490 to 523°C) based on twofeldspar thermometry and reduce (low \(fO_2\)) nature of NG and TG plutons are consistent with the above suggested conditions for tin mineralization. However, there is need of mineral exploration based detailed investigation of TG and NG, which may help locating the probable hidden tin deposits in the study area.

7.5 Geochemical Petrology

7.5.1 Granite typology and protolith assessment

Since past four decades, granites types are popularly classified alphabetically (I-, S-, M-, H-types etc.) based on their type of sources involved in the production of felsic melts. The alphabetic classification scheme I-type and S-type of granitoid respectively derived from igneous and sedimentary sources was proposed by Chappell and White (1974) from Lachlan fold belt of Australia, which correspond to metaluminous (molar A/CNK <1.05) and peraluminous (molar A/CNK >1.05) types respectively according to alumina saturation index (Shand, 1943). However, S-type, peraluminous granitoids can be produced by partial melting of sources other than sedimentary protoliths (Miller, 1985). Other granite-type such as A-type can be characterized as alkaline, anhydrous and anorogenic related granites formed in extensional environment (Loiselle and Wones, 1979). Progressively other alphabetic terms came into existence which are imprecise, such as C-type for charnockitic granites (Kilpatrick and Ellis, 1992); G-type for granitic origin (Wang et al., 1991), H-type for hybrid granitoids (Castro et. al., 1991); M-type mantle origin in an island arc tectonic setting (White, 1979). However, among these types I, S and A-type granites became most popular for characterizing the source of felsic melts.

The stock-like granitoid plutons exposed within the Mahakoshal Belt are broadly classified as metaluminous (I-type) to peraluminous (S-type) granitoids (Fig. 5.1a) based on their alumina-saturated index (Shand, 1943) and mineral assemblages (Chappell and White, 1974). Metaluminous I-type includes JG and HG plutons, whereas intermediate between metaluminous (I-type) to peraluminous (S-type) nature is KG pluton, and rest of the plutons (DG, RG, NG and TG) are peraluminous (S-type) in nature. The
metaluminous to peraluminous nature of KG could be either resulted from gradual fractionation or involvement of mantle and crustal derived components in their genesis. However, it is noteworthy that peraluminous (S-type) KG do not contain any aluminous mineral phase typical of crustal signature (e.g. garnet cordierite, sillimanite, muscovite) as commonly found in peraluminous variety of RG (garnet) and two-mica NG and TG (muscovite). It is therefore inferred that peraluminous KG cannot be derived from sedimentary protolith rather fractionation of mafic-minerals (pyx-hbl-bt-pl-ttn-mag) might have played important role to generate peraluminous S-type of residual melt from dioritic type of KG parental magma. Typical crustal origin for NG, TG and RG plutons can be suggested because of their exclusive and highly peraluminous (S-type) nature and presence of aluminous mineral assemblage such as garnet and muscovite (e.g Turpin et al., 1990; Patiño Douce, 1991, 1999). Moreover, RG, NG and TG are devoid of microgranular enclave but leucocratic peraluminous variety of KG contains hybrid microgranular enclaves. NG, TG and RG do contain broken detrital and inherited zircons, which strongly suggest involvement of sedimentary protoliths in their genesis. Two-mica NG and TG do not contain garnet but garnet is present in RG which is devoid of muscovite. However, secondary laths of sericite type of muscovite can be noticed within plagioclase of RG formed due to later thermal effect. Garnet is also present in highly metaluminous (I-type) HG pluton but it is of magmatic origin inferred from wavy contact with plagioclase and also biotite crystallized within fracture system of early phase garnet.

Modified alkali-lime index (MALI, Frost et al., 2001; Frost and Frost, 2008) have suggested calc-alkaline (DG, TG), alkali-clacic (KG, RG, NG, HG), and alkaline (JG) series of granitoid plutons (Fig. 5.2c), which are not always consistent with observed mineral assemblages and other geochemical parameters. However, these series dictate that they mostly evolved through fractional differentiation mechanism except to those do not follow a linear or curvilinear trends. Therefore, KE, RE, JE samples plot in more than one series because of modification by some processes such as magma-mixing and mingling other than fractional crystallization.

Based on whole rock Nb-Y-Ce content anorogenic granites can be subdivided into A1 and A2 groups (Eby, 1992) corresponding to sources like ocean island basalt, derived by crystal fractionation or partial melting, and continental crust source (excluding metasediments) or derived from partial melting of arc-type sources, respectively (Eby, 1992). JG pluton exclusively belongs to A-1 type whereas other granitoids correspond to
A-2 type including the syn-collisional two-mica NG and TG (Fig. 7.9). JG represent

**Fig.7.9** Nb-Y-Ce Triangular plot (after Eby, 1992) for various granitoid plutons of Dudhi Gneissic Complex, Mahakoshal Belt. A-1: apparently related to source of lower crust derived from partial melting or fractional crystallization in within plate tectonic setting. A-2 granitoids derived by melting of continental crust in syn-to-post tectonic environment.

A-2 granitoid plutons (HG, KG, DG, and RG) appear as products of post-orogenic processes, formed by melting of continental crust and/or arc-type mafic magma sources. Mafic magma sources most likely signify the contribution of mafic to hybrid magma input to the granitoids partly modifying the original composition.
of granitoids. However, in terms of Ga-Al parameters (Whalen, 1987) most granitoids occupy in I and S-type fields and gradually graded into A-type, whereas some of them such as JG and HG represent exclusively A-type granitoids (Fig. 7.10 a, b). It is therefore likely that granitoids plutons may represent latest manifestation of the processes related to the crustal delamination by mafic magma underplating of a thickened continental crust. It has triggered extensional-related, post-collisional magmatism which induced generation of metaluminous (I-type) to partly A-type granitoid plutons.

A-type Proterozoic post-collisional granitoids in the world were probably formed in oxidizing conditions which were derived from melting of quartz-feldspathic igneous sources (Anderson and Morrison, 2005; Dall’Agnol and Oliver, 2007; Zhao and Zhau, 2009; Zhang, et al., 2011). This optimism may support the view that the granitoids of Mahakoshal Belt were more likely generated by partial melting of Archean continental crust (Bundelkhand craton). It is consistent with field and petrographic features of basement xenolithic-body of Archean (Bundelkhand) component engulfed into the volcanics of Mahakoshal Belt. A-type granites generally require high temperature melting (Clemens et al., 1986) which can be achieved by underplating of mafic magma from lithospheric mantle upwelling in extensional tectonic regime causing decompressional melting (e.g. Castro et al., 2012). In collisional related tectonic setting lowering of melting point of crustal rocks may have triggered by crustal thickening with low conductivity heat loss (Beaumont et al., 2004).

Multicationic R₁-R₂ diagram (Fig. 5.3b) of DeLa Roche (1980) takes into account of almost all major cations, which can be used more confidently to characterize individual rock and rocks in association. Most of the studied rocks are silica-saturated belonging to monzodiorite → tonalite → granodiorite → granite series (KG) and syenodiorite →quartzmonzonite → granite series (JG). JE belongs to nephline syenite, silica-undersaturated rocks because of high K₂O and mafic nature. The NG, TG and RG belong to typical granodiorite to granite series. These igneous series should have been formed in a wide tectonic regimes ranging from syn-collision to post collisional, even to rift-related environment (e.g. Bachelor and Bowden, 1985).

Continental crustal may produce compositionally diverse nature of magmas under variable melting conditions such as pressure, temperature, fugacities (fO₂, fH₂O) (e.g. Wolf and Wyllie, 1994; Patino Douce, 1996; Patino Douce and McCarthy, 1988; Singh
Compositional similarities or dissimilarities can be produced by partial melting of continental crust initiated by underplating of mafic magma which may or may not be involved in mixing process. It can be visualized on the basis of major oxides ratios in which the compositions of granitoids from various plutons of Dudhi gneissic complex, Mahakoshal Belt have been compared with the experimental melt compositions (Fig. 7.11). Partial melting of metabasalt and amphibolites might have involved for the generation of RG melt which grades towards melt compositions formed by melting of metagreywacke protolith. It supports the field and petrographic observations. RG clearly intrude the volcano-sedimentary lithounits manifested by swarms of metavolcanics xenoliths engulfed into it. It is more probable that hybrid sources (sedimentary and metabasalt) with dominant metavolcanics could have been partially melted to produce RG melt. Two-mica, peraluminous NG and TG occupy the metagreywacke field and one sample plots into melt produced by melting of pelitic rock. Thus the NG and TG plutons should have been exclusively formed by melting of metagreywacke with small input of pelitic type of sedimentary protoliths, which is equivocally supported by the presence of abundant inherited and detrital zircons. However, granitoids of HG and JG plutons and enclaves do not provide clues for any protolith as they do not plot in any of melt protolith fields. It is likely that they may have genetic connotation with mantle-related sources. It is supported by the presence of mafic
microgranular enclave (JE) in JG but is absent in highly metaluminous HG which is hybridized by the extensive assimilation of deeper-derived hbl-diorite xenolith hosted in it. It may also be inferred that these melts (JG and HG) represent derivatives of mantle-derived mafic magma of diverse compositions. Aplitic veins associated with JG and HG plutons suggest participation of metagreywacke protolith but aplitic melts represent extreme fractionated residual melts from JG and HG parental magmas, and hence enriched in K₂O and Na₂O contents. Thus, melt composition alone cannot be used for source characterization of aplitic melts. KG samples plot in a wide range of melt composition fields showing curvilinear negative trend, which should have resulted from fractional differentiation of KG parental melt (that could be fractionated from mantle-derived because diorite and mesocratic variety of KG melt do not plot in any of the melt composition field). However, in the generation of the mesocratic to leucocratic KG and also DG melts mixed (mantle and crustal) sources were involved. It is possible that with the progressive fractionation of mantle-derived melts assimilation of overlying continental crust (meta-volcanics and meta-granite) has taken place (e.g. Clemen, 2012).

It is because of fact that the observed mineral assemblages do not support the assimilation crustal rocks and therefore do not contain peritectic mineral assemblages (garnet, sillimanite, cordierite and muscovites, but contain mafic-hybrid microgranular enclaves even in leucocratic granite of KG.

### 7.5.2. Depth of granitoid magma emplacement based on normative composition

JG samples plot towards Or-end and progressively evolved towards eutectic point at 1 kbar in the ternary Ab-Or-Qz system (Fig. 7.12). However, JG do not represent eutectic composition at 1 kbar most likely because of some chemical modification and high K₂O content. JG melt might have emplaced at shallow level as evident from Al-in-amphibole barometer. HG also occupy the primary phase field near Or-corner, suggesting saturation of HG melt in Or-component, similar to as noted for JG magma, at deeper level equivalent to P_H₂O of 5-10 kbar. Least silica KG melt (diorite) plots in Ab-primary phase field, then grade stowards Or-Qz cotectic. The KG samples do not display linear array and no eutectic-like composition (Fig. 7.12), which cannot be solely accounted by crystallization rather mixing-mingling processes between mafic and felsic end-members (e.g. Dokuz, 2011) and also by assimilation noted in leucocratic variety of KG. Melanocratic to mesocratic KG melt probably crystallized at 5-10 kbar. DG melt may have crystallized between 5 and10 kbar but orthoclase dominates over albite. The
estimated solidification pressures for granitoid plutons of Dudhi gneissic complex, Mahakoshal Belt, are consistent with the Al-in-amphibole barometers. Amphiboles, are however, absent in highly evolved peraluminous, leucocratic granitoids (NG, TG and RG), the whole-rock normative composition of which can be used to predict solidification pressures using Ab-Or-Qz ternary systems. The NG, TG, RG and JG melts plot between 1 and 5 kbar eutectics which indicate shallow level emplacement, and rest of other plutons were emplaced at still deeper levels. Shallow level emplacement of (NG, TG, RG and JG) plutons is also corroborated by the presence of country-rock xenoliths such as metapelite chert and metavolcanics, except in TG which has rare or no xenolith because of its windy nature of intrusion. Plutons of shallower levels that emplaced at 1-3 kbar are considered to be low-temperature granites (e.g Johannes and Holtz, 1996) and they plot to the right of cotectic/eutectic minima of the H2O-saturation (metaluminous-JG; peraluminous-NG, TG, RG plutons). Other than these, diorite of KG melt plots left to the cotectic/eutectic minima of the H2O-saturation and initially melt was anhydrous during crystallization of KG. The fugacity of water (fH2O) has increased which resulted crystallization of amphibole and biotite at relatively shallow level. The HG and DG also crystallized at deeper levels.
7.6 Recognition of magmatic processes

7.6.1 Geochemical evolution of HG and HE

7.6.1.1 Mantle derived HG melt fractionation without assimilation

Alkaline, magnesian and highly metaluminous nature of monzodiorite HE and quartz-monzodiorite HG plot at low-silica end and HG show evolved fractionated trend on Harker’s variation diagrams (Fig. 5.7). Hbl-diorite enclave (xenolith HE) contains higher amount of MgO, FeO, CaO, MnO, P₂O₅ as compared to host HG, which suggest relatively more mafic nature of HE than HG. Moderate degree of fractionation involving plagioclase, hornblende, magnetite sphene, apatite, and zircon assemblage has caused decreasing contents of MgO, Fe₂O₃, MnO, CaO, Al₂O₃, P₂O₅ with increasing SiO₂ (Fig. 5.7). K-feldspar might have not precipitated in early stage of crystallization as it behaves as incompatible elements with increasing degree of fractionation. HE samples do not exhibit fractionation trend as they cluster at lowest silica end, but may represent cumulate formed by early minerals segregation of HG magma (e.g. Dahlquist, 2002). The likely cumulate origin of HE has been discarded on the firm ground that HE lacks typical cumulate texture and exhibits medium grained granular texture (Fig. 4.2e), and moreover cannot represents fine grained border facies of pluton as HE is present as bigger xenolithic block in HG pluton (Fig. 2.4f). However, both the HE and HG bear some signatures of mantle components in their genesis as reflected by the presence of olivine, diopside and hypersthene norms. Relatively mafic and felsic nature of HE and HG can be explained by the dominance of An-norm and Or-norm respectively.

Relatively higher content of Sc, V, Cr, Cu, Zn, Sr, Y and lower content of Rb, Ba, Th, U, Pb (Fig. 5.11a) as compared to those of other granitoids suggest least or no crustal contamination during evolution of HG. Mantle-normalized trace element patterns of HG and HE exhibit low degree of positive Ba and U anomaly, which indicate role of metasomatized mantle whereas observed negative Th, Nb, Ti, Eu, Sr anomalies suggest low degree fractionation of mantle-derived magma. Low degree of negative Y-anomaly of HG (Fig. 5.11a) is most probably caused by amphibole and garnet fractionation from HG melt. Low negative Y-anomaly in HE (hbl-diorite xenolith) could be retention of garnet in residue. Plagioclase fractionation is manifested by decreasing Al₂O₃, CaO with increasing SiO₂ and negative Eu-anomaly (Fig. 5.11b). Both HE and HG exhibit identical trace and REE patterns, which most likely reflect involvement of same protolith in their genesis but at different time because of xenolithic (older) nature of HE than HG.
Fig. 7.13 SiO₂ vs Major oxides (wt%) and trace element (ppm) variation diagrams showing linear to sub-linear trends for JG (open circle) while diorite of JG (filled square) and JE (filled circles) plot mostly off the granitoid trends forming a distinct compositional group. JE and host JG are joined by tielines (after Bora, et al., 2013). Note the high K₂O content of microgranular enclaves, except one, as compared to the host granitoids.
Alternatively, HE may represent unmelted protolith from which HG magma has derived. Aplitic vein cross-cutting the HG shows inclined convex LREE and concave HREE patterns with pronounced negative Eu-anomaly, and strong positive anomalies for Rb, Th, La Ce, and highly depleted Nb, Ti, Sr, Eu, transitional elements, which together suggest more evolved residual phase of HG parental magma.

7.6.2 Geochemical evolution of JG pluton

7.6.2.1 Mixing between fractionating mafic and felsic magmas

Diorites are enriched in compatible elements compared to granitoids of JG pluton, which may indicate either their co-magmatic origin or formation by combined mixing-fractionation processes. On most Harker’s diagrams (Fig. 7.13), broadly two distinct compositional groups of studied rocks can be recognized. One is formed by granitoids and the second, distinct, by diorite and ME. Granitoids define negative linear to curvilinear evolutionary trends for MgO, TiO$_2$, CaO, Fe$_2$O$_3$, MnO, P$_2$O$_5$ and, to some extent, Al$_2$O$_3$ and K$_2$O, with increasing silica, which strongly suggests fractional crystallization of a hbl+cpx-bt-pl-kf-mag-ap-ttn assemblage. However, Na$_2$O shows a positive trend indicating that sodic mineral such as albite-rich plagioclase did not participate in the fractionating assemblage. Most ME and diorite plot off the major-element trend of granitoids except for TiO$_2$, Fe$_2$O$_3$, and MnO where diorites follow the trend of granitoids at low silica end. P$_2$O$_5$ contents of two samples, one each from diorite and ME, roughly follow the trend of granitoids as did the CaO of diorites and a ME. Compositional contrasts between ME and host granitoids are indicated by tie-lines (Fig. 7.13). It is surprising to observe that the ME being more basic than granitoids bear higher K$_2$O contents than their respective host granitoids. Granitoids form clearly defined negative linear trends for Ni, Cr, V, Sc, Zn and $\sum$REE and, to some extent, also for Sr, Ba, U, Th, Zr, and Nb with increasing silica. These are consistent with the behavior of major elements, which are commonly substituted by counter trace elements. In the case of Zn and Sc, diorites and ME lie on the low silica-end of the compositional trend of granitoids. Cu, Ga, Nb and Y (not shown) exhibit moderate degree of scatter for the granitoids, and even for diorites and ME. Although, diorites and ME do not show apparently co-linear variations against silica, they together form a separate compositional group on most of the binary plot (Fig. 7.13), unrelated to the trends of granitoids. However, it is important to note that diorites and one ME are substantially enriched in
MgO, CaO, MnO, Ni, and depleted in Al₂O₃, K₂O as compared to the rest of the ME samples.

Evolution of some key elements (e.g. MgO, Sr, Cr) against silica (Fig. 7.14) can demonstrate significant genetic role to recognize the operative processes in the temporal and chemical evolution of diorites, JE and granitoids of JG plutons. Diorite containing least silica (~50 wt%) and enriched MgO, Sr and Cr may represent a mafic magma end-member which evolved most likely along steeply declining trends of these elements (shown by broken arrows), which progressively mixed with fractionating granitoids (shown by large ellipse) forming the hybrid JE at various stages (shown by solid arrows) of syn-crystallization with increasing polymerization and crystal load (e.g. Barbarin, 2005; Slaby and Martin, 2008). It is however likely that the non-linear trends for some trace elements could have been caused by non-linear (chaotic) mixing phenomena (e.g. Perugini et al., 2008; Slaby et al., 2011). Position of another diorite (quartz diorite) with respect to granitoids suggests that it could be an early hybridization product between pristine diorite (mafic) and least fractionated granitoid melts, consistent with field observations. The two compositional groups of studied JG rocks may thus represent mixing between syn-crystallizing (or fractionating) mafic and felsic magmas.

### 7.6.2.2 Chemically modified JE

Out of four studied JE, three JE are peraluminous (molar A/CNK = 1.16–1.21), nepheline-corundum normative, and poor in CaO-Sr despite of their low SiO₂ and high MgO-Fe₂O₃ contents. These geochemical features strongly suggest that the JE globules in JG were chemically modified by crustal component and xenocryst incorporation from
host granitoids during mixing event causing selective enrichment in $\text{Al}_2\text{O}_3$, Rb and Ga. In these JE An-rich part of plagioclase is highly resorbed and sericitized, which could have depleted the ME in CaO and Sr and enrich them in $\text{K}_2\text{O}$ and Rb (e.g. Dorais et al., 1990; Eberz and Nicholls, 1990). Tie-lines joining the JE-host granitoid on $\text{Na}_2\text{O}$, $\text{K}_2\text{O}$ vs $\text{SiO}_2$ diagrams (Fig. 7.14) suggest that $\text{K}_2\text{O}$ migration from felsic host to the JE was more rapid than that of $\text{Na}_2\text{O}$ during co-mingling and strong undercooling of JE. This is consistent with experimental results (e.g. Watson and Jurewicz, 1984) and the observations made elsewhere (e.g. Kumar et al., 2004b; Kumar and Pieru, 2010). Indeed the quenching of JE would only allow the most rapid processes such as exchange of alkalies to have significant effect across the JE-host granitoid boundary (e.g. Wiebe et al., 1997). Positive Rb anomaly with negative Ba and Sr anomalies on mantle-normalized trace-element patterns of JE and JG (Fig. 5.12a) suggest that they are dominated by crustal component. However, Sc, V, Zr and Cu contents close to the normalizing values may indicate mantle contribution in their sources (e.g. Silva et al., 2000). Involvement of varying proportions of crustal components in the mantle-derived JE might explain the observed compositional range of JE from metaluminous to peraluminous types, which should have been accelerated when the JE magma interacted with the crustally-derived JG melt. The JE as small-volume melts once injected into the granitic host became a more viscous, forming discrete magma globules within the chamber, and quickly reached thermal equilibration with host magma long before chemical equilibrium was achieved (Waigh et al., 2001 and references therein). The JE globules may then have cooled at a relatively slower rate allowing considerable chemical exchange between JE and host magma. This process should have led to the formation of medium- to even coarse-grained JE with mixed crustal and mantle-related chemical signatures. Mingling of hybrid JE into host granitoids indicates that magma-mixing between coeval mafic (diorite) and felsic melts must have occurred at depth (hidden hybrid zone) prior to its injection into the host magma (Bora et al., 2013).

7.6.2.3 REE features of mafic and felsic magma fractionation

The chondrite-normalized REE patterns of diorites-JE and JG are compatible with a moderate degree of fractionation of a mafic melt similar to the diorite (without Eu anomaly) and felsic parental melt respectively, which upon fractionations gave rise to residual melts with high sum of REE and negative Eu anomalies (Fig. 5.12a, b). However, fractionated cumulate rock-types with positive Eu anomalies have not been
recorded. The JE have lower sum of REE compared to the host JG and show almost identical REE patterns with negative Eu-anomalies similar to those for JG, which strongly opposes the cumulate (cognate) or restite origin for JE but favours their interactions synchronous with fractional crystallization.

### 7.6.2.4 Evidence of chemical equilibration

![Spidergrams and rare earth element patterns](image)

**Fig. 7.15** Primitive Mantle-normalized (Taylor and McLennan, 1985) trace element spidergrams chondrite-normalized (Taylor and McLennan, 1985) rare earth element patterns plotted for microgranular enclave (filled circle) and respective host granitoid (open circle).

Primitive mantle-normalized trace and chondrite-normalized REE patterns of JE and a JG pairs (Fig. 7.15) are identical and parallel, and even a pair of AE1 – AG5 overlaps each other suggesting partial to complete trace-element (including REE) equilibration. It
could have been due to varying degrees of elemental diffusion across the mafic-felsic interface primarily controlled by the size of the JE and its resident time (e.g. Kumar and Rino, 2006 and references therein). Rapid diffusional transfer of K and H$_2$O (plus vapour) promotes conversion of pyroxene into amphibole and biotite (Waight et al., 2001) as observed in some JE of JG which also acted as a major carrier of elements through vapour transport during JE-host JG magma interaction prior to complete solidification of the system.

**Fig. 7.16a-h** JG felsic magma crystallization and fractionation with increasing viscosity and K$_2$O enrichment in the residual JG melt along with formation of various types of microgranular enclave by different nature of mafic magma injection and resulted in enclave formation through convection.

Based on the geochemical feature of mafic to hybrid enclave (JE) and their host JG (Fig. 7.16a-h), it is evident that there was existence of hybrid zone beneath JG melt.
formed as a result of complete mixing between mafic and felsic melts. This hybrid zone is injected and disaggregated as JE globules (Fig. 7.16c, d), which contains K-feldspar xenocrysts and equilibrated chemically with the host JG pluton since it has more residential time of cooling and thus hybrid JE became coarse grained in textural feature. With progressive degree of crystallization of JG melt another pulse of mafic magma injected (Fig. 7.16e, f), which may have resulted in the formation of coarse grained xenocryst-free JE that partially equilibrated with host JG melt. Finally at high degree of fractionation of JG melt, again mafic magma injected (Fig. 7.16g, h) which quenched or undercooled against relatively cooler JG melt when the adjacent residual melt was enriched in $K_2O$ content and thus $K_2O$ rapidly diffused into mafic JE (Watson and Jurewicz, 1984; Kumar and Rino, 2006) but they are partly or less equilibrated with host JG melt. The hybrid JE globules are bigger in size than those of other JE. It is more likely that total consumption of hybrid zone occurred before the injection of another pulse of mafic magmas with increasing degree of fractionation of JG host melt. Further, low degree of mixing occurred in the later phase of JE formation because the host JG melt had attained high viscosity and as a consequence, rate of diffusion of other elements ceased (except for $K_2O$) into the mafic JE magma globules, which are fine grained.

7.6.3 Geochemical evolution of KG pluton

On Harker’s diagrams (Fig. 5.7) MgO, Fe$_2$O$_3$, MnO, CaO, Al$_2$O$_3$, Na$_2$O, TiO$_2$ of KG behave as compatible elements because of crystallizing ferromagnesian (Hbl-bt) and pl-qz minerals, which are consistent with modal mineral variations among the KG samples. Diorite and enclave (KE) are enriched in MgO, Fe$_2$O$_3$, MnO, CaO, TiO$_2$ content with least silica content suggesting their mafic nature, which interacted with felsic host KG melt synchronous with fractionation. Na$_2$O and K$_2$O behave as incompatibe elements because of least role of K-feldspar in the evolution of KG magma.

Trace elements such as Sc, V, Cr, Ni, Cu, Zn are enriched in KE (enclave) and diorite because of their mafic nature, and are therefore depleted in large ion lithophile elements. On Harker’s plots (Fig. 5.8) Rb, Ba, Pb, Th, U contents of KG show positive correlation with silica suggesting less or no role of minerals hosting these elements at least at an early stage of KG melt evolution. However, Zr content in KG is slightly scattered but broadly show positive correlation with SiO$_2$ which could be because of zircon sorting during KG fractionation.

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Diorite and enclave (KE) contain dominantly mantle-derived component as reflected from diopside and hypersthene norms and absence of corundum norm. However, KG other than diorite and KE, display corundum, diopside and orthoclase norms, which suggest recycling of infra-crustal continental sources in the generation of KG melt.

7.6.3.1 Elemental exchange during mingling of mafic-felsic magmas

![Diagram](image.png)

**Fig. 7.17a-d** Primitive mantle-normalized (Taylor and McLennan, 1985) trace element and chondrite-normalized (Taylor and McLennan, 1985) rare earth element-patterns for KE (opened circle) and respective host KG (open circle) pairs.

Microgranular enclaves (KE) are hosted in melanocratic and leucocratic variety of KG, and normalized trace and rare earth element patterns of KE-KG pairs are plotted (Fig. 7.17a-d) in order to infer the chemical relationships between them. KE hosted in leucocratic KG have shown distinct elemental compositions, except Rb, U, Nb, La, Y, and Zr which are almost at the same level most likely caused by selective elemental diffusion at varying rate. If this is true then K should have also been equilibrated between them because of high diffusion coefficient. Fast undercooling of KE globules into KG host might have also inhibited the elemental exchanges between them (Kumar and Pieru, 2010) but undercooling does not support through petrographic feature of KE as it is coarse grained and contains cylindrical apatite crystals (Fig. 4.5d). It is most probable that less temperature difference between mafic and felsic magmas may have been repulsive.
factor to form acicular apatite in KE. It is therefore suggested that chemical feature of KE in leucocratic KG should have been affected by incorporation of xenocrystic phase (biotite, zircon) during mixing with partially crystalline host KG. Higher sum of REE of KE than the KG do not support cumulate origin for KE. On the other hand, KE hosted in melanocratic KG has shown sub-parallel to parallel trace and REE patterns, except for Ba, Ti, Eu, which were most likely controlled by varying degrees diffusion of elements between them during mingling of KE globules before the entire system solidified (e.g. Kumar and Rino, 2006). Thus the resident time of semi-solid KE was sufficiently longer to facilitate the elemental exchanges.

**7.6.3.2 Fractional crystallization (FC): evidence from REE chemistry**

The variation in sum of REE and moderately inclined REE patterns with varying degree of negative europium anomalies (Fig. 5.13b) in KG are related with fractional differentiation of KG melt primarily involving hbl-bt-pl assemblage, and therefore leucocratic KG exhibits high degree of negative Eu-anomaly. KG with flat REE, least SiO₂ without any Eu-anomaly may represent primary KG melt which fractionated to form mesocratic to leucocratic varieties of KG. Mantle-normalized spidergram suggests role of both mantle- and crustal-derived components in the generation of KG melt which subsequently experienced fractional crystallization. Interestingly diorite and melan-to-mesocratic KG show flat and negative Th-anomalies respectively, which suggest release of Th and U as incompatible elements into the lower continental crust during melting of mantle-related rocks and residual remained in depleted mantle (Hofmann, 1997). Thus the diorite might have formed directly from the mantle-related rocks. The observed Th positive anomaly in leucocratic KG suggests that radioactive minerals precipitated in the residual felsic melt rather than being accommodated in meso-to-melanocratic variety of KG.

**7.6.4 Geochemical evolution of DG pluton**

Although only three samples of DG are analysed for major and trace including rare earth elements, major oxides (MgO, MnO, TiO₂, P₂O₅, Fe₂O₃, CaO and Al₂O₃) and trace elements (Sc, V, Cr, Ni, Cu, Zn, Sr, Y, ΣREE) are found decreasing with increasing SiO₂ on Harker’s plots (Fig. 5.14a), which indicate fractionation of hbl-bt-pl-mag-ilm-ap assemblage, also supported by observed negative Eu-anomaly with higher sum of REE in
evolved DG sample (D8). Other two samples do not exhibit Eu-anomaly, and most likely they represent composition close to primary DG melt (Fig. 5.14b).

7.6.5 Geochemical evolution of RG pluton

Since only three RG samples have been investigated, therefore process diagnosis such as fractional differentiation cannot be inferred to a high level of confidence. Nevertheless, some broader petrogenetic interpretation can be made. Microgranular enclave (RE) in RG have relatively higher MgO, Fe₂O₃, MnO than those host RG (Fig. 5.7), which suggest mafic nature of enclave RE melt. Modal abundance of apatite-needles in RE suggests P₂O₅ saturated RE melt (Fig. 5.7), which quenched against relatively cooler host RG to form acicular apatite (e.g. Kumar, 1995). Apatite has also observed in host RG but they are stubby in nature and lower in amount than RE. Y content in RG is higher as compared to other granitoids, except NG and TG, and exhibits positive correlation with silica, which is because of peritectic garnet and monazite particularly in more evolved RG melt. Ba shows negative correlation with silica but K₂O is positively correlated with silica (Fig. 5.7), because of preferential partitioning of Ba in early K-feldspar crystal depleting Ba quicker than K₂O in evolving melt. Hence Ba decreased with increasing K₂O in the RG residual melt as similarly argued elsewhere (e.g. Heier and Taylor, 1959; Berlin and Henderson, 1969, Kerrick, 1969; Higgins and Kawachi, 1977; Pierozynski and Henderson, 1978; Michael, 1981; Mehnert and Büsch, 1981).

7.6.5.1 Evidence of trace and REE partial equilibration

Trace and rare earth elements of two RE-RG pairs are plotted (Fig. 7.18a-d), which are identical for most of the elements but a relatively higher degree of negative anomaly is noted for Ba, U, Sr, Eu and Ti in RG as compared to that of RE. Differences in degree of negative anomaly for these elements can be very well explained in terms of differential degrees of fractionation of respective melts involving some phases (Kf-pl, accessory minerals) hosting these elements, which strongly suggest that RE and RG have experienced mixing mingling and partial chemical equilibration during their internal fractional differentiation mechanism bringing them to parallel to sub-parallel elemental levels (e.g. Kumar and Rino, 2006). Moreover, identical patterns of RE and RG can never be explained to their derivation from a common source because of mafic nature of RE than the host RG.
Geochemical Evolution of NG and TG plutons

7.6.6.1 Protolith for generation of leucocratic NG and TG melts

In the Harkar’s variation diagrams (Fig. 5.7), both NG and TG clustered at high silica end because of leucocratic nature, absence of Fe-Mg and lower content of accessory phases. They contain high content of alkalies and low amount of MgO, CaO, Fe₂O₃, MnO and P₂O₅. Trace elements such as transitional and high field strength elements (HFSE) are depleted but large-ion lithophile elements (LILF) particularly Rb and Y are enriched because of their derivation from crustal protoliths and role of monazite and/or xenotime during melting events respectively. Rb/Sr ratio is extremely high for NG (Rb/Sr = 9.04-3.11) suggesting its typical crustal origin because Sr-preferentially retained in plagioclase residue. Sum of the REE is low in both NG and TG which is a characteristic feature of leucocratic melt derived from crustal sources. Dominance of corundum norm and absence of olivine, diopside and hypersthene norms are indicative of pure crustal-derived melt without any lower-crustal or mantle input.
7.6.6.2 Low-to-high degree of partial melting

NG exhibit higher degree of LREE to HREE fractionation with high degree of negative Eu-anomaly as compared to those of TG (table 5.6; Fig. 5.16b), which can be
explained by low degree of melting of metagreywacke source, which can produce NG melt and at relatively higher degree of melting of same protolith can form TG parental melt. However, plagioclase fractionation may have played important role for observed negative Eu-anomaly. It is however likely that source was already depleted in plagioclase. Evolved phase of NG melt (NT-1) is highly depleted in ∑REE with slightly inclined REE pattern. The observed mantle-normalized negative anomaly for Ba, Nb, Ti and pronounced positive anomaly for Rb and Y (Fig. 5.16a) are characteristics of crustal origin. Role of xenotime and/or monazite signifies the observed positive Y anomaly.

7.6.7 Comparison of lithounits with lower, upper and bulk continental crusts

Average trace and rare earth elemental compositions of diorites, granitoids of various studied plutons and primary nature of volcanics have been compared with calculated compositions of lower, upper and bulk continental crusts (Taylor and McLennan, 1985), which are shown in figures 7.19a-f. This exercise has been carried out assuming that average whole rock composition of granitoids for each studied pluton may represent the composition of parental melt. Average K, Rb, Pb, Ba, Th, U, Sr contents of JG are higher than those of lower, upper and bulk crusts, which suggest evolved nature of JG melt with more crustal input in its genesis. This is further supported by depletion in transition metals as compared to crustal rocks (Fig. 7.19a). Broadly KG, DG, HG and to some extent JG melts have shown trace element patterns similar to upper crust, and are evolved in nature in terms of incompatible (large ion lithophile) elements but are highly depleted in transition metals, which signify involvement of crustal components in their genesis. Evolved nature of JG, KG, DG and HG melts is very much reflected by elevated LREE patterns and moderate degree of negative Eu-anomalies as compared to those of crustal rocks (Fig. 7.19b). It should be noted that there are not much differences in the HREE patterns of lower, upper and bulk continental crusts. K, Rb, Pb, Th, U contents of RG, NG, TG are higher than the crustal compositions, but transition metals (Sc, V, Zn, Cu, Ni, Cr) are highly depleted which are indicative of pure crustal origin for these melts(Fig. 7.19c). RG melt is highly evolved in terms of LREE and negative Eu-anomaly whereas LREE patterns of NG and TG melts are similar to those of upper and bulk continental crust respectively, which are indeed reflection of differences in crustal source rocks in their genesis (Fig. 7.19d) or could be related to differential degrees of partial melting of the same crustal source, as discussed earlier. In terms of incompatible trace elemental patterns, diorite of JG has resemblance with upper continental crust whereas
Fig. 7.19 c, d Comparison of lower, upper and bulk continental crust compositions (Taylor and MacLennan, 1985) with (c-d) average composition of individual peraluminous plutons; RG, NG and TG.
KG diorite relates to bulk crust (Fig. 7.19e). Volcanics representing primary melts have, however, shown elemental patterns similar to bulk and lower crusts, which are suggestive of relatively mafic nature of volcanism with addition of crustal input (K, Rb) at least in

Fig. 7.19 e, f Comparison of lower, upper and bulk continental crust compositions (Taylor and MacLennan, 1985) with (e-f) primary volcanics melts (R-10 and R-12) diorites (JG, KG)
the case of one volcanic melt (Fig. 7.19e). LREE pattern of JG diorite is more evolved than upper crust, which suggests crustal involvement in the genesis of JG melt whereas LREE pattern of KG diorite is similar to bulk crust pointing to lesser amount of crustal component in KG diorite as compared to JG diorite (Fig. 7.19f). The average compositions of studied lithounits from various plutons of Dudhi gneissic complex suggest involvement of more crustal components in their genesis with a minor mafic input similar to the lower crustal composition.

7.6.8 Tectonic setting of plutonic and volcanic lithounits

7.6.8.1 Syn-to-post collisional felsic plutonism

![Tectonic discrimination diagrams](image)

Fig. 7.20a-b Tectonic discrimination diagrams (after, Pearce, et al., 1984) based on Nb, Y and Rb suggest formation of most granitoid plutons post-collisional (dotted circle) tectonic setting.

The granitoids plutons of Dudhi gneissic complex from Mahakoshal Belt largely bear mineralogical and geochemical characters similar to syn-to-post collisional granitoids, which could have been evolved in divergent or intraplate regime (e.g. Roberts and Clemens, 1993; von Blanckenburg et al., 1998) However, some plutons (JG, HG) have shown some features similar to high-K calc-alkaline granitoids, which appear to have formed in convergent tectonic setting (e.g. DePaolo, 1981; Hildreth and Moorbath, 1988, Barbarin, 1990); where partial melting of mantle triggered by dehydration of subducted slab enriched in fluids may be contaminated by continental crust during ascent through mantle wedge. Interestingly tectonic discrimination diagrams (Fig. 7.20a, b) based on Nb, Y and Rb suggest formation of most granitoid plutons in post-collisional tectonic setting. Undoubtedly two-mica NG and TG were emplaced in syn-collisional
tectonic setting, which was closely followed by post-collisional plutonism during crustal relaxation stimulated due to mafic magma underplating beneath thick-skinned crust of Mahakoshal Belt. The likely mechanism is that the lower crust was heated up by juxtaposed asthenosphere that caused partial melting of lower crust and thus generated high-K, calc-alkaline metaluminous (I-type) felsic magmatism, that is another tectonic scenario for generation of calc-alkline type of magmatism (Houseman, et al., 1981, Moyen et al., 2003), which does not required Himalayan-type subduction (classical-arc) related orogeny (e.g. Li et al., 2002a, Zhang et al., 2007).

Some post-collisional granitoids (JG, HG) have affinity with metaluminous A-type granitoids as per the geochemical criteria of Whalen et al. (1987) and Eby (1992). The granitoid plutons of metaluminous A-type, high-K calc-alkaline characters have generally been reported not only in anorogenic rift environment but also in post-collisional tectonic setting (Collins, 1982, Sylverster, 1998; Barbarin, 1999). Most of the high-K calc-alkaline post-collision types of magmatism are also reported from other parts of the Paleoproterozoic orogeny belts world-wide e.g. Limpopo Belt in Zimbabwe, Trans Northern China, Trans Amazonian orogeny in South America, Capricone orogeny in West Australia, Paleoproterozoic orogeny of Baltic, Greenland, Siberia, North America, South Korea, and Central India Tectonic Zone (Zhao, et al., 2004; Beakhouse and Davis, 2005; Li et al., 2007).

7.6.8.2 Fractional crystallization assimilation (AFC) in post-collisional volcanics

The volcanic lithounits have fractionating trend on Harker’s variation diagrams (Fig. 5.9), which demonstrate negatively correlation linear trend for MnO, MgO, Fe₂O₃, CaO against SiO₂. Crustal contamination through assimilation synchronous with fractionation has been evidently shown by scattered data points of Al₂O₃, Na₂O, K₂O with increasing silica, as equivocally suggested by field and petrographic observations. P₂O₅ content in volcanics is increasing initially with increasing silica upto medium silica level and then P₂O₅ inflected decreasing with increasing silica, which is because of entry of apatite crystal in crystallizing sequence at middle stage of fractionation. Volcanics with low silica content are enriched in TiO₂, Fe₂O₃, MgO, MnO, CaO, Sc, V, Cr, Ni, Cu, Zn, Nb (Fig. 5.10) and depleted in in other large ions lithophile elements, which probably represent primary feature (subalkaline, tholeiitic; Fig. 5.5d) of volcanics from
Mahakoshal Belt. Assimilative nature of melt can be recognized by higher Rb content attained at intermediate level of silica with progressive precipitation of K-feldspar as the melt evolve. U, Th, Pb behaves as incompatible element because of less or no role of radioactive minerals during crystallization, and hence enriched in residual melt. Low Zr containing volcanics most likely represent least crustally-contaminated samples occurring in the inner zone of the volcanic sequence whereas high Zr volcanics have assimilated...
crustal rocks representing strongly contaminated volcanic rocks. It is thus inferred that with the progress of differentiation volcanics have experienced crustal-contamination.

Least silica containing volcanics show flat to slightly inclined REE patterns without any significant Eu-anomalies, which may represent tholeiitic to mildly alkaline nature of primary mafic melts and have subsequently fractionated with assimilation while

![Fig. 7.23 Sr/Y-Y diagram (Drummond, 1990) shows island arc affinity for volcanics and dykes.](image1)

![Fig. 7.24 Ce/Nb vs. Th/Nb plot (after Song et. al., 2004 in Xiaoqi et. al., 2009) showing the bimodal distribution of volcanics and mafic dykes.](image2)
eruption (e.g. DePaolo, 1981). These primary mafic melts are characterized by low LILE and high HFSE contents (Fig. 5.17a) whereas other volcanic samples are products of different degree of fractionation and assimilation reflected in positive Rb, Ba, Th and negative Sr, Eu, Nb, Ti anomalies (Fig. 5.17a). The volcanic rocks of such chemical features can be achieved in an volcanic arc system due to partial melting of lithospheric mantle by subducting slab or in an intraplate rift-related tectonic setting (e.g. Long et al., 2006) or in post-collisional setting (e.g. Zhao, 2007; Xiaoqi, 2009). These volcanic rocks indeed belong to intraplate basalts, however two samples may represent primary mafic melts showing MORB type affinity characterized by low Zr and Zr/Y ratio (Fig. 7.21), which made them distinct from arc volcanics originated in subduction setting. Arc related volcanism formed by partial melting of subducting slab is higher in Nb content with low La/Nb ratio and mostly associated with adakitic melt (Sajona et al., 1994; Castillo et al., 2002). However, Nb-rich basalt can also occur in post-subduction (Castillo et al., 2008) and post-collisional tectonic settings (Guo, et al., 2007). The absence of adakite-type rock suite and island arc type features of studied volcanics (Fig. 7.22) (because of contamination) indeed point to post-collisional nature of volcanism in Mahakoshal Belt.

Most of the dykes and two volcanic samples (primary mafic melt) plot in Nb-enriched basalts field whereas rest of the volcanics lie in island arc basalt having low P2O5 and high P2O5 characters (Fig. 7.23), which may be due to contamination with Neoarchean crustal components (Bundelkhand craton, the xenoliths of which are found hosted in volcanics). The volcanics have passed through the basement lithology in extension regime and have got contaminated. This hypothesis is further supported by the observed higher Ce/Nb and Th/Nb ratios (Fig. 7.24), which most likely resulted from assimilation of subduction-related Neoarchean crustal component (e.g. Song et al., 2004) during mafic magma underplating. Primary mafic melts corresponding to OIB (overlapping with CFB), have escaped contamination, which should have been derived from primitive mantle. Field association and penecontemporaneous nature of volcano-sedimentary succession have unambiguously indicated extensional-tectonic environment of deposition.

7.7 U-Pb SHRIMP zircon chronology: inferred sequence of geological events

Zircons from studied mafic-felsic lithounits are magmatic in origin except a few rims of zircons from the DG pluton, which have recorded the event of metamorphism at
ca 1000 Ma. NG and TG plutons contain heterogeneous population of inherited zircons derived from recycling of Neoarchean crust. Overall U-Pb SHRIMP zircon chronology of studied lithounits (tables 6.1-6.7; Figs. 6.1-6.24) suggest occurrence of bimodal magmatic episodes and number of geological events in the evolution of Dudhi gneissic complex of Mahakoshal Belt, Central India Tectonic zone.

7.7.1 Opening of the Mahakoshal Basin: evidence from inherited and magmatic zircons of NG and TG plutons

The opening of Mahakoshal Basin on the Neoarchean basement (Bundelkhand craton) was first and foremost stage of evolution of Mahakoshal Belt in which sediments, mostly derived from the Neoarchean early earth crust, were deposited. This is manifested by the old inherited zircon cores hosted in two-mica NG pluton, which yielded heterogeneous $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 2832, 2511, 2491, 2451 and 2092 Ma, as remnants from sedimentary sources as detrital zircons. A wide range of zircon ages (2832-2092 Ma) strongly suggests heterogeneous nature of sedimentary deposition which dictate minimum age of ca 2.0 Ga for the evolution of Mahakoshal Basin, or in other words, basin was opened ca 2.0 Ga. Most of the inherited cores of NG zircons are well rounded with magmatic zones, which suggest zircons were eroded out from igneous provenances, and deposited as detrital zircons in sedimentary basin. The basin was initially opened during sagging of the Bundelkhand craton forming intracratonic type of basin as consequence of mafic-magma underplating. The mantle-derived mafic magmas upwelling resulted in upliftment of Bundelkhand craton and initiated volcanism forming the Bijawar and Gwalior Basins in the rift-related tectonic environment (Sharma and Rahman, 1996) almost synchronous with the evolution of an intracratonic basin of Mahakoshal Belt. Based on Rb-Sr whole rock dating an age of ca 1830±200 Ma for the volcanics of Gwalior Group has been suggested, which erupted in basinal margins of Bundelkhand craton (Crawford, 1970; Crawford and Compston, 1971). Bundelkhand craton was also intruded by a number of episodic mafic dyke swarms during 2150-2000 Ma (Rao et al., 2000), which were synchronous with sedimentation in the adjacent intracraonic basin of Mahakoshal Belt. These dykes have undergone deformational event at ca 1800 Ma (Rao et al., 2005), which most likely represents closing time of the Mahakoshal Basin.
7.7.2 Slow rate of Mahakoshal Basin closure: evidence from collisional-related plutonism

The first collisional event was marked by generation of NG pluton at ca 1880 Ma, which triggered closing of the Mahakoshal Basin. It could have been initiated due to the amalgamation of northern Bundelkhand craton with the southern Bastar craton. The collision of cratons could have been driven due to the continuous insistent of Bundelkhand craton towards south by rift-related volcanism in the Bijawar and Gwalior Basins. At the same time probably southern (opposite part of the plate) margin of the Bundelkhand craton subducted beneath Bastar craton forming calc-alkaline Cu-(Mo-Au) hosting Malanjkhand granitoid plutons. This was perhaps the first Palaeoproterozoic subduction which occurred in the southern part of the Bundelkhand craton, not in the northernmost part of Mahakoshal Belt supporting the subduction model of Yedekar (1990) Yedekar et al. (2003) and Stein, et al. (2004). Thus, there was synchronous coupled-tectonic activity (subduction and extension at opposite ends of the Bundelkhand craton) was operated in Bundelkhand craton where northernmost part was experiencing mafic magmatism in Bijawar and Gwalior in extensional regime and evolution of Mahakoshal Belt on the other southern end of subduction. It also suggests a complete Wilson cycle operated in Bundelkhand craton at boundary between Neoarchean and Paleoproterozoic.

The collisional event was continued at late or final stage of closure of Mahakoshal Basin, which has taken at least ~100 Ma time span to close completely at ca 1780 Ma manifested by formation of two-mica bearing TG pluton. It is also inferred that the collision-related compression was almost in N-S direction whereas minimum stress should have been perpendicular to the compressional tectonic forces.

7.7.3 Bimodal mafic-felsic magmatism in post-collision tectonic setting

The least stress was operating in the E-W direction, which induced eruption, emplacement and easy passage for the vast coeval bimodal mafic-felsic magmatism of Mahakoshal Belt in post-collisional tectonic setting. It is mainly recorded by mafic to porphyritic volcanism and felsic to intermediate plutonism (JG, JE and DG) occurred at ca 1750 Ma, which postdates the collisional events after a gap of ca 30 Ma. Further thickening of crust occurred as a result of collision and E-W directed minimum stress perpendicular to the strike of compressional stress, and thickened crust delaminated
followed by decompressional melting during crustal relaxation. It led to advancement of extensional regime in Mahakoshal Belt, which prompted volcanic eruption at 1746±9.3 Ma rift-like (sensu lato) environment, which is almost synchronous to felsic plutonic activities (~1750 Ma) having post-collisional tectonic affinity. These regions were highly affected by decompressional melting events and the melts formed at initial stage were rich in H₂O, CO₂ and halogens, which lowered the temperatures of the magmas (McKenzie, 1989). This is also because of fact that extensional tectonics serve as potential conduits for the transportation of volatile-rich fluids in overlying lithosphere through the crystallizing magma under subsolidus condition (e.g. Da Silva Filho, 1993) consistent with the obtained subsolidus temperature (~450-650°C) using two-feldspar thermometry of granitoid plutons of Mahakoshal Belt. The prevailing physical conditions of magmas and tectonic environment may actually be responsible for sporadic or hidden ore deposits in the study area.

7.7.4 Coeval mafic-felsic magma interaction in plutonic setting

U-Pb SHRIMP zircon chronology of mafic-to-hybrid JE and KE and host JG and KG magmas underlines their coeval nature. The obtained ²⁰⁷Pb/²⁰⁶Pb mean ages of 1756.8±6.5 Ma and 1745±16 Ma are zircon crystallization age in metaluminous and peraluminous JE respectively, which are similar to the zircon crystallization ²⁰⁷Pb/²⁰⁶Pb age of 1752±4 Ma of host JG magma within the limit of associated errors. Thus, these three magmas of diverse compositions co-existed and interacted (mingled and mixed) in plutonic condition forming the bulk of JG pluton. Weighted mean ²⁰⁷Pb/²⁰⁶Pb age 1717.3±7.1 Ma of zircons in KE and 1729.2±7.2 Ma age of zircons in KG strongly suggest their coeval nature that also mingled and mixed in plutonic conditions. However, the ²⁰⁷Pb/²⁰⁷Pb age of zircon crystallization (1737±9 Ma) in HE is synchronous with KG (~1729 Ma), older than the host HG (1713±5 Ma), and is therefore HE is considered xenoliths caught up by HG magma.

7.7.5 Intra-plutonic sedimentation

The obtained ²⁰⁷Pb/²⁰⁶Pb zircon ages from quartzite, in which HG pluton has intruded, suggest existence of two provenances of ca 1729.2±6.3 Ma and 1763±7 Ma which fed the sediments for the deposition of quartzite. These provenances were most likely represented by Katoli pluton and vast mafic-felsic lithounits of the study area. Based on the obtained ages, the geological correlations of the events can be speculated
that a small basin was opened after 1730 Ma felsic magmatism (KG) and most of the sediment were dominantly derived from Katoli pluton and nearby regions, which might have exhumed or eroded insitu and deposited sedimentary lithofacies. The deposition of sediments was essentially ceased before the intrusion of HG pluton (1713±5 Ma).

7.8 A viable model for the evolution of Mahakoshal Belt

The E-W trending Central Indian tectonic zone (CITZ; Radhakrishna 1989 and Acharyya and Roy, 2000) divides the Peninsular India into northern Bundelkhand craton and southern Bastar, Dharwar, Singhbhum cratons. The CITZ is considered as a suture between northern Bundelkhand and Southern Bastar craton (Bhowmik et al., 1999, 2000; Acharyya, 2003; Mall et al., 2008; Majumder and Mantani, 2009; Nagananjeyulu and Santosh, 2010). The Bundelkhand and Bastar cratons are composed of granite gneiss-supracrustal belts, which recorded the thermal history of Archean era from ca 3.5 to 2.5 Ga (Basu, 1986; Sarkar et al., 1993, 1995; Sharma, 1998; Roy et al., 2000). The Mahakoshal Belt represents northernmost part of CITZ, which is bounded by SONA fault system in north and south that is considered different tectonic components of an asymmetrical rift basin (Roy and Bandyopadhyaya, 1990 a,b), intracratonic basin (Yedekar et al., 1990) or intraplate-graben (Srivastava et al., 2000). Nair et al., (1995) provided the stratigraphy of Mahakoshal Belt, which was later modified by Roy and Devarajan (2000). They opined that the Mahakoshal Belt is a pericraton formed along the marginal part of Bundelkhand craton in a shallow marine environment and followed by rift-related tectonic setting (Chaudhuri and Basu, 1990; Kumar, 1993; Nair et al., 1995) in which alternate sequence of volcano-sedimentary deposits took place (Fig. 7.3). However, timing of rifting and formation of rift-related back arc or intracratonic basin and sedimentation is still a matter of debate.

The supracrustal rocks of Mahakoshal Belt have experienced deformation which formed upright folds to slightly overturn fold in ENE-NSW striking and axial planes dipping (Roy and Prasad, 2003). Supracrustal rocks have undergone low grade metamorphism (Roy et al., 2002) as a consequence of felsic intrusive magmatism at ca 1.8-1.5 Ga (Sarkar, et al., 1998). The basement for Mahakoshal Belt is Bundelkhand craton, which yielded an age of ca 2.2 Ga (Sarkar, et al., 1995) commonly exposed to the northern fringe of Mahakoshal belt (Fig. 2.2). Xenolithic block of the granite basement is also observed within mafic-ultramafic volcanics of Mahakoshal Belt of Sidhi region (Figs
2.10e, f and 7.1a-c), which way be equivalent to Neoarchean granitoids of Bundelkhand craton. The eastward extension of Mahakoshal Belt is considered Chotanagpur Gneissic Complex (CGC; Sen, 1956; Baidya et al., 1989; Ray Barman et al., 1990) which evolved at 1.6-1.5 (Acharyya, 2003 and references therein). It has further continuation into northeastern part of India i.e. Shillong-Meghalaya Plateau (Evans, 1964) and has continuity with Eastern Ghat Mobile Belt (EGMB) in southwestern India (Fermor, 1936) forming S-shaped sigmoidal Proterozoic mobile belts, which are relics or reworked zones of Archean components (Rickers et al., 2001; Ramachandra and Roy, 2001).

Based on whole rock Rb-Sr (Ray Barman et al., 1990) and U-Pb zircon chronology, age of metamorphism at ca 1.6-1.5 ca of intrusive charnockites of CGC with later tectonic imprint at ca 1000 Ma has been envisaged (Ray Barman et al., 2002 in Acharyya, 2003). The similar timing of metamorphism and tectonism is also recorded in the present study by U-Pb SHRIMP zircon chronology of Dudhi gneissic complex, Mahakoshal Belt. Felsic magmatism occurring at ca 1.5 Ga has been correlated with continental-continental collision of the Bundelkhand craton with Bastar craton (Roy and Prasad, 2003). However, the obtained chronological data on TG (~1.9 Ga) and NG (~1.8 Ga) of Dudhi gneissic complex of Mahakoshal Belt do not support the age of collision at ~ 1.5 Ga rather suggests minimum age (~1.9-1.8 Ga) of collision at ca 1.9 Ga, which at least continued ca 100 Ma duration as recorded from the ages of two-mica granitoid plutons. These collisional-related granitoid plutons (NG and TG) were generated during closing of Mahakoshal Basin. Detrital inherited zircon cores (2.8-2.0 Ga) are hosted in NG magma that strongly suggest involvement of heterogeneous Neoarchean provenances in the sedimentation of basin of Mahakoshal Belt, that formed prior to 2.0 Ga and closed before the intrusion of ~1.9 Ga NG pluton during continental-continental collision (assembly) of Bundelkhand craton with Bastar craton. The collision-related orogeny led to production of thickened continental crust. Consequent to the formation of thick continental crust, the Mahakoshal Belt experienced a period of relaxation in the form of extension and delamination of lower crust simultaneously. The delamination of lower crust was caused by mafic magma underplating beneath the Neoarchean continental crust, also evident from geophysical investigation that the Mahakoshal Belt is underlain by thick pile of high density mafic-ultramafic rocks and bounded by deep seated fault (Kaila, 1988). Mafic magma underplating caused decompression melting and thermal erosion from the base of the crust and adiabatically rising of melts which may have involvement
of different proportions of mantle and crustal components. These are represented by vast post-collisional volcano-plutonic, high-K calc-alkaline magmatism at ca 1.75 Ga in the evolution of Dudhi gneissic complex of Mahakoshal Belt. Bimodal mafic to hybrid (enclaves) and felsic magmatism (JG and DG) are coeval and interacted and emplacement at shallower to mid-crustal levels. Earlier the mega thermal event at ~1.75 Ga was correlated with the closing of the rifted Mahakoshal Basin (Acharyya, 2003), which is not per se but actually post-collisional, extensional regime. However, post-collisional tectonic regime lies between plate-collision and intraplate setting (Liégeois, 1988) generally known as mature orogeny (e.g. Rolland, 2012). During this period highly oxidized felsic magmatism took place, recorded from JG pluton, less oxidised DG melt, though both are coeval magmatic bodies. A difference in oxygen fugacity ($f_{O_2}$) is determined by their emplacement level or type of source regions. Thus, the Mahakoshal Belt is devoid of subduction-related magmatic bodies and truly related to syn-to-post collisional bimodal (mafic-felsic) magmatism that have played major role in the formation of continental crust of Mahakoshal Belt. Hence, it opposes the northward subduction tectonic model as suggested elsewhere (e.g. Bhowmik et al., 1999; Roy and Hanuma Prasad, 2001a; Roy and Prasad, 2003). More viable model of Mahakoshal Belt is southward subduction of southern margin of Bundelkhand craton beneath Bastar craton, which probably produced calc-alkaline Manjkhkhand pluton (Yedekar, 1990) and followed by the formation of Mahakoshal Belt which was at the evolutionary stage of intracratonic basin.

### 7.9 Comparison of past Mahakoshal orogeny with present younger Himalayan orogeny in Indian subcontinent

The Mahakoshal Belt neither represents a back-arc nor fore-arc basins. This is because of fact that in a back-arc basin synchronous volcano-sedimentary sequences should have been deposited prior to collision-related magmatism, which may be produced during closing of a basin as observed in Shyok suture zone of Himalayan orogeny. Fore-arc basin should be associated with subduction-related (arc) magmatism, which is not recognized in Mahakoshal Belt. Since oldest magmatism in Mahakoshal Belt is collisional-related granitoids closely followed by post-collisional bimodal mafic-felsic volcano-plutonic magmatism, thus it represents an intraplate basin formed within Bundelkhand craton, which may be a final product of metacratonization (e.g. Liégeois, et al. 2012). Post collisional magmatism can however be associated with orogenic belt related to subduction magmatism after the collision but their characteristic products are in
the series of subduction (e.g. Ladakh batholith)-collision-(two-mica granitoids of Ladakh pluton)-post-collision-(Karakoram batholith) magmatic activities. Therefore, initiation of the Mahakoshal Belt is manifested by sedimentation (opening of a basin) intracratonic progressively followed by closing event forming collisional-related NG and TG plutonism and terminated with post-collisional, metaluminous to peraluminous granitoid plutonism during crustal relaxation (JG, KG, DG, HG, RG and volcanic lithounits). Later
the entire Dudhi gneissic complex of Mahakoshal Belt has undergone shearing tectonism at ca 1000 Ma (Fig. 7.25).

![Diagram](image)

**Fig. 7.26** Columbian supercontinent assembly at ca 1700 Ma recorded in various orogens of the continents including the Indian subcontinent (after Bose et al., 2011; Goodge et al., 2001), which is also recognized in the Mahakoshal Belt located in the north of Central Indian Tectonic Zone (CITZ).

Accretionary history of Columbian supercontinent (ca 1760–1600 Ma) has been recorded globally prior to Rodinia at ca 2100–1800 Ma (Rogers and Santosh, 2002; Zhao et al., 2004; Bose et al., 2011), whereas the felsic magmatism occurred at about 1750–1700 Ma in parts of Laurentia and Baltica (Rogers and Santosh, 2009; Bose et al., 2011 and references therein). The obtained SHRIMP zircon ages of NG and TG suggest ca 1900-18000 Ma collision event, followed by vast mafic-felsic magmatism ca 1750 Ma of JG, DG, and volcanic lithounits along with coeval mafic and felsic magmatism and their interactions in respective plutons of Dudhi gneissic complex in Mahakoshal Belt of Central Indian Tectonic Zone (CITZ) during the period of Columbian crustal consolidation (Fig. 7.26). Their vestiges are also reported in Aravalli (e.g. Bhowmik et al., 2010, Kaur et al., 2013) and Eastern Ghat Mobile Belt (Bose et al., 2011) of Indian cratonic regions.

The Paleo-to-Mesoproterozoic, high to low-K, metaluminous to peraluminous (I-to-S type) intrusive and extrusive mafic-felsic magmatism, in the northernmost part of CITZ i.e. Mahakoshal Belt are products of *Metacratonization*. These granites and volcanics are inter-linked in space and time and some of them have gradational to intrusive
relationships with each other. Most of the I-type granitoids have more or less similar mineralogical, major and trace elemental chemistry and few of them contain coeval mafic to hybrid microgranular enclaves. Magma-mixing, fractional crystallization, differential degree of melting and assimilation have been the major processes operated in syn-to-post collisional tectonic settings.

The granitoid bodies mostly cooled below subsolidus conditions, which determine the role of hydrothermal fluids and fluorine or halogens emanated from granitic system of Mahakoshal Belt. The country rocks are highly contaminated with fluorine, which interacted and mixed up with groundwater available to local people of the Sonbhadra area, for drinking purpose. It is more hazardous and causes major bones and dental diseases (mottled enamel) in human being (Zipkin, et al., 1958) that was complained by local people during geological field investigations.