An overview on geochemistry of Proterozoic massif-type anorthosites and associated rocks

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A critical study of 311 published WR chemical analyses, isotopic and mineral chemistry of anorthosites and associated rocks from eight Proterozoic massif anorthosite complexes of India, North America and Norway indicates marked similarities in mineralogy and chemistry among similar rock types. The anorthosite and mafic-leucomafic rocks (e.g., leuconorite, leucogabbro, leucotroctolite, anorthositic gabbro, gabbroic anorthosite, etc.) constituting the major part of the massifs are characterized by higher $Na_2O + K_2O$, Al_2O_3, SiO_2, Mg# and Sr contents, low in plagioclase incompatible elements and REE with positive Eu anomalies. Their $\delta^{18}O‰$ (5.7–7.5), initial $^{87}Sr/^{86}Sr$ (0.7034–0.7066) and ε_{Nd} values (+1.14 to +5.5) suggest a depleted mantle origin. The Fe-rich dioritic rocks occurring at the margin of massifs have isotopic, chemical and mineral composition more close to anorthosite–mafic-leucomafic rocks. However, there is a gradual decrease in plagioclase content, An content of plagioclase and X_{Mg} of orthopyroxene, and an increase in mafic silicates, oxide minerals content, plagioclase incompatible elements and REE from anorthosite–mafic-leucomafic rocks to Fe-rich dioritic rocks. The Fe-rich dioritic rocks are interpreted as residual melt from mantle derived high-Al gabbro melt, which produced the anorthosite and mafic-leucomafic rocks. Mineralogically and chemically, the K-rich felsic rocks are distinct from anorthosite–mafic-leucomafic-Fe-rich dioritic suite. They have higher $\delta^{18}O$ values (6.8–10.8‰) and initial $^{87}Sr/^{86}Sr$ (0.7067–0.7104). By contrast, the K-rich felsic suites are products of melting of crustal precursors.

1. Introduction

Excellent reviews summarizing various geological aspects including field setting, petrology and geochemistry of the massif anorthosite complexes are provided by Emslie (1973, 1985), Duchesne and Demaiffe (1978), Bose (1979), Morse (1982), Ashwal (1982a, 1993), Duchesne (1984), Duchesne *et al* (1985), Leelanandam (1987), Wiebe (1992), and Leelanandam and Narsimha Reddy (1998). The significant inferences made in these reviews include that the anorthosite massifs of dominantly Proterozoic age (1.7–0.9 Ga) consists of anorthosite, leucogabbro, leuconorite and/or leucotroctolite. They are often associated with Fe-rich dioritic rocks at the margin and a suite of K-rich plutonic rocks ranging in composition from granite, granodiorite to monzonite. The massifs are variably deformed and unmetamorphosed to metamorphosed in subgreenschist to granulite facies conditions (Ashwal 1993). In anorthositic rocks, plagioclase ($An_{\sim 50\pm 10}$) is the dominant mineral with subordinate amounts of orthopyroxene, clinopyroxene and/or olivine. They are characteristically high in Al_2O_3, CaO, Na_2O, Sr, LREE and Eu with positive Eu anomaly of variable intensities and low in other oxides and trace elements (Ashwal 1993). Most commonly, the felsic plutonic rocks are reportedly younger than the massif and represent crustally derived anatectic melts, broadly coeveal, but genetically unrelated to the anorthosites (Ashwal and Seifert 1980;

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Duchesne *et al* 1985; McLelland *et al* 2004). By contrast, Bose (1979) and Demaiffe and Hertogen (1981) argue that the massif anorthosite and the bordering granitoids suite are comagmatic. Fe-rich dioritic rocks with higher concentration of Ti and P typically occur as thick sheets, veins, veinlets and apophyses at the interface between the anorthosite and the felsic rocks. The ferrodiorites are variably interpreted to be:

- residual melts of deep crustal fractionation of tholeiitic magma that produced the anorthosite (Ashwal and Seifert 1980; Ashwal 1982b; Weibe 1990; McLelland *et al* 1994; Mitchell *et al* 1996),
- melts generated through deep crustal anatexis of basic to intermediate rocks (Duchesne *et al* 1985), and
- immiscible liquids from mangeritic magma having no genetic link to anorthosite (Owens and Dymek 1992).

Researchers who consider anorthosite–leuconorite–ferrodiorite and granitoids to constitute a comagmatic suite propose lower crustal melts (monzodiorite and even granodiorite) as possible parental magmas (cf. Demaiffe and Hartogen 1981). Others who consider the felsic rocks to be genetically unrelated to the anorthosite–leuconorite–ferrodiorite suite propose fractionated basic magmas as parental to the anorthosite suite (Simmons and Hanson 1978; Mitchell *et al* 1995). Mitchell *et al* (1995) have argued that high-Al gabbros (HAG) characterized by high Al_2O_3 (15–19 wt.%), enriched REE patterns with positive Eu anomaly, low initial $^{87}Sr/^{86}Sr$ (0.7033) and high ε_{Nd} (up to +2) are ideally suited to be magmas parental to the Laramie anorthosite.

The massifs of Eastern Ghats Mobile Belts (EGMB, India) and Rogaland (SW Norway) have marked similarities with those of the Greenville Belt of North America (Berthelsen 1980; Leelanandam 1987). However, in spite of noted similarities among the anorthosite complexes, published work, which compare up-to-date information on the geochemistry and geology of the massif anorthosites and associated rocks involving Indian massif anorthosites is lacking. In many complexes, Fe-rich dioritic rocks and HAG, the most important key units for genetic interpretations are known after the existing reviews. Moreover, the salient findings of the existing reviews suggest that a general consensus is far from being attained regarding the magmatic relationship among various members of the anorthosite massifs and of the anorthositic rocks with the bordering K-rich felsic suite. Disputes also prevail regarding nature of anorthosite parental magmas. Thus, the present study mainly aims at (i) reviewing the state-of-the-art knowledge on whole rock geochemistry along with isotope chemistry, mineral chemistry and general geology of the Mid-Proterozoic massif-type anorthositic and associated K-rich felsic rocks for a general comparison in order to bring out their commonalities and differences, and (ii) to explain the observed chemical characteristics in the light of magmatic relationship among various members and anorthosite parental magmas.

With this end in view, a minimum of eight representative complexes such as Bolangir, Chilka Lake, Jugsaipatna and Turkel (EGMB, India); Marcy (known as 'typical' of Middle Proterozoic massifs worldwide) and Morin (Grenville province, North America); Nain (Nain province, North America); and Hidra (Rogaland complex, SW Norway) are taken into consideration (figure 1). The basic information used in the present overview are collected from the published literature, for example, Raith *et al* (1997), Bhattacharya *et al* (1998), Dobmeier (2006) and Nasipuri and Bhattacharya (2007) on Bolangir complex; Sarkar *et al* (1981), Mukherjee *et al* (1999), Raith *et al* (2007), Chatterjee *et al* (2008) on Chilka complex; Nanda and Panda (1999), and Joshi *et al* (2006) on Jugsaipatna complex; Maji (1997), Maji *et al* (1997) and Maji and Sarkar (2004) on Turkel complex; De Waard (1970), Seifert *et al* (1977), Simmons and Hanson (1978), Ashwal and Seifert (1980), Ashwal and Wooden (1983), Morrison and Valley (1988), Ashwal (1993) and McLelland *et al* (2004) on Marcy massif; Papezik (1965), Heath and Fairbairn (1969), Barton and Doigs (1972), Martignole (1974) and Seifert *et al* (1977) on Morin massif; De Waard and Wheeler (1971), Wiebe (1978), Xue and Morse (1993), Emslie *et al* (1994), and Bedard (2001) on Nain massif; Demaiffe and Hertogen (1981), Duchesne *et al* (1985), Demaiffe *et al* (1986), and Demaiffe and Javoy (1980) on Hidra massif.

2. General geology

As most of the anorthosite bodies referred in this study have been mapped and described in several publications, only a brief description focusing the important relevant points are given here. The mineral chemical data and isotopic chemistry of different rocks are summarized in table 1.

2.1 *Eastern Ghats massif anorthosite complexes*

Out of the nine known massif anorthosite occurrences in the Eastern Ghats Mobile Belt (EGMB), Bolangir (∼400 km^2), Chilka Lake (∼250 km^2),

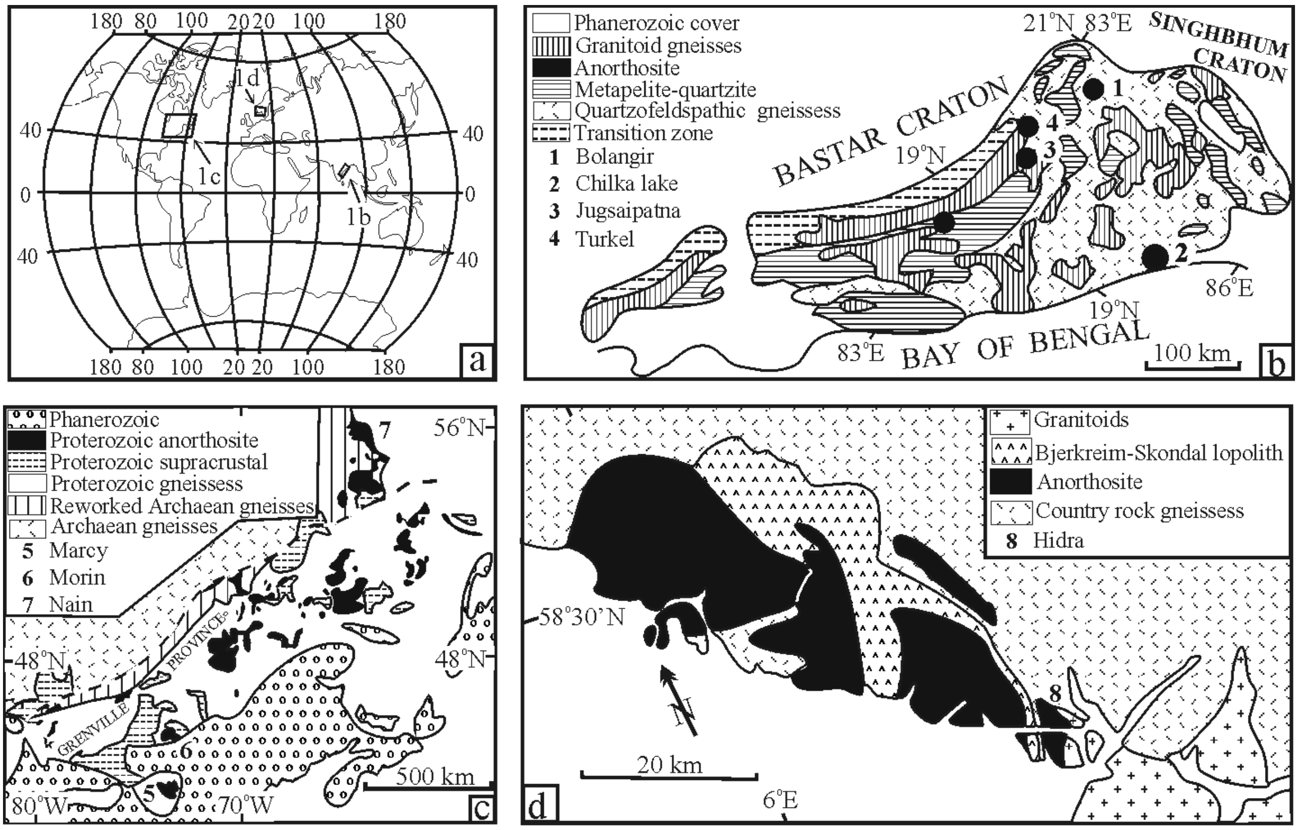

Figure 1. **(a)** Index map showing the location of Eastern Ghats mobile belts, India; Eastern Canadian shield, America; and Rogaland complex, Norway. **(b)** Generalized geological map of Eastern Ghats mobile belt showing the locations of major massif anorthosite complexes (modified after Leelanandam and Narsihma Reddy 1998; Ramakrishnan *et al* 1998). **(c)** Generalized geological map of Eastern Canadian shield showing the locations of major massif anorthosite complexes (modified after Ashwal 1993), and **(d)** Generalized geological map of Rogaland igneous complex, SW Norway showing anorthosite complexes (modified after Duchesne *et al* 1985).

Jugsaipatna ($16\,km^2$) and Turkel ($81\,km^2$) massifs have been studied in detail by several researchers over the last few decades (figure 1b). The age of anorthosite magmatism in EGMB varies widely as ca. 1400 Ma (Sarkar *et al* 1981), 792 ± 2 Ma (Krause *et al* 1998) and 983 ± 2.5 Ma (Chatterjee *et al* 2008) for Chilka Lake complex and ca. 930 Ma (Krause *et al* 1998) for Bolangir complex. The massifs are hosted by a migmatitic garnetiferous felsic suite of rocks varying in composition from granite to monzonite through granodiorite. They contain variable proportion of perthitic K-feldspar>antiperthitic plagioclase, quartz, orthopyroxene, clinopyroxene, pre- to syn-kinematic porphroblastic garnet, ilmenite, apatite, zircon, rutile, hornblende and biotite. The garnet commonly encloses subrounded grains of quartz, K-feldspar, biotite, plagioclase, ilmenite, zircon and apatite suggesting their formation by incongruent melting (Bhattacharya *et al* 1998). Within the massifs, broadly a zonal distribution with homophanous, coarse grained anorthosite (*sensu stricto*) in the central portion and leuconorite characterized by crude phase layering in the marginal part is obvious. Anorthosites and leuconorites are coarse-grained, leucocratic rocks containing plagioclase, orthopyroxene $\gg$ clinopyroxene, apatite, ilmenite, magnetite, biotite and K-feldspar as primary minerals and biotite, hornblende, muscovite, calcite, and coronal garnet as secondary minerals. Melanocratic fine-grained ferrodioritic rocks (also variably named by several researchers as jotunite in Chilka Lake and gabbronorite in Jugsaipatna complexes) occur as fairly continuous bands, discontinuous lenses and pods at the immediate contact with the massifs. Ferrodioritic rocks having blastoporphyritic texture defined by antiperthitic plagioclase phenocrysts set in a well recrystallized fine-grained mosaic of Fe-rich orthopyroxene$\gg$Fe-rich clinopyroxene, ilmenite with or without fayalitic olivine. The proportion of ferromagnesian minerals increases from anorthosite to ferrodiorite through leuconorites. The ferrodiorite shares intrusive relationship with the massifs ranging from a few centimetres to a few tens of metre wide dykes crosscutting banding of leuconorites (Maji *et al* 1997; Raith *et al* 1997; Bhattacharya *et al* 1998;

Table 1. *Summary of mineral chemical data and isotopic chemistry of different lithounits in the massif anorthosite complexes.*

	Lithounits	Pl (X_{An})	Opx (X_{Mg})	Cpx (X_{Mg})	$\delta^{18}O$‰	Initial $^{87}Sr/^{86}Sr$ (age Ma)	ε_{Nd}
B	Anorthosite	0.55–0.67	0.55–0.60	0.68–0.75	6.7–7.5[a]	–	–
	Mafic-leucomafic rocks	0.50	0.37–0.55	0.60–0.66	6.8–7.4[a]	–	–
	Fe-rich dioritic rocks	0.40–0.48	0.17–0.42	0.30–0.53	7.1–8.7[a]	–	–
	Felsic rocks	–	–	–	7.5–9.2[a]	–	–
CL	Anorthosite	0.64–0.70	0.65–0.70	–	–	0.7066 (1404)[b]	–
	Mafic-leucomafic rocks	0.50–0.70	0.55–0.65	–	–	-do-	–
	Fe-rich dioritic rocks	0.45–0.48	–	–	–	–	–
	Felsic rocks	0.41–0.48	–	–	–	0.7067 (1312)[b]	–
J	Anorthosite	0.60–0.70	0.40–0.61	–	–	–	–
	Mafic-leucomafic rocks	0.70	0.37	–	–	–	–
T	Anorthosite	0.48–0.52	0.42–0.49	0.54	7.0–7.3[c]	–	–
	Mafic-leucomafic rocks	0.41–0.43	–	0.37	6.9–7.2[c]	–	–
	Fe-rich dioritic rocks	0.39–0.47	0.27	0.34–0.41	6.9[c]	–	–
	Felsic rocks	0.26–0.45	0.01–0.50	0.14–0.46	8.0–10.8[c]	–	–
Ma	Anorthosite	0.43–0.55	0.68–0.78	–	3.0–9.5[d]	0.7040–0.7042 (1290)[e]	+1.14[e]
	Mafic-leucomafic rocks	-do-	-do-	–	-do-	0.7039–0.7041 (1290)[e]	+1.85 to +4.08[e]
	Fe-rich dioritic rocks	0.34–0.38	0.51	0.55	–	0.7042–0.7045 (1290)[e]	+0.44 to +1.93[e]
Mo	Anorthosite	0.47–0.53	0.60–0.80	–	–	0.7040–0.7063[f,g]	–
	Mafic-leucomafic rocks	-do-	-do-	–	–	-do-	–
	Fe-rich dioritic rocks	0.45–0.48	0.64	–	–	–	–
	Felsic rocks	0.25–0.35	0.38	–	–	0.7050 (1008)[f,g]	–
N	Anorthosite	0.47–0.52	0.58–0.68	0.51		0.7034–0.7052 (1300)[h]	−3.0 to −13.9[h]
	Mafic-leucomafic rocks	0.43	0.46–0.51	–	–	-do-	-do-
	Fe-rich dioritic rocks	0.35–0.45	0.30–0.35	–	–	0.7056–0.7090 (1300)[h]	−8.1 to −11.7[h]
	Felsic rocks	0.15–0.40	–	–	–	0.7061–0.7104 (1300)[h]	−8.7 to −14.1[h]
H	Anorthosite	0.47–0.54	–	–	5.7[j]	0.7057 (1000)[i]	+5.5[i]
	Mafic-leucomafic rocks	0.42–0.49	0.60–0.66	–	5.2–6.4[j]	0.7055 (1000)[i]	+4.7[i]
	Fe-rich dioritic rocks	0.47	–	–	5.2[j]	0.7061 (1000)[i]	+2.7[i]
	Felsic rocks	–	–	–	6.8–7.4[j]	0.7086 (1000)[i]	+1.6[i]

The mineral abbreviations are after Kretz (1983). B = Bolangir, CL = Chilka Lake, J = Jugsaipatna, T = Turkel, Ma = Marcy, Mo = Morin, N = Nain, H = Hidra. (Data sources: [a]Bhattacharya *et al* 1998, [b]Sarkar *et al* 1981, [c]Maji and Sarkar 2004, [d]Morrison and Valley 1988, [e]Ashwal and Wooden 1983, [f]Heath and Fairbairn 1969, [g]Barton and Doig 1972, [h]Emslie *et al* 1994, [i]Demaiffe *et al* 1986, Demaiffe and Javoy 1980).

Dobmeier 2006; Nasipuri and Bhattacharya 2007). In Bolangir, Chilka Lake and Jugsaipatna complexes, HAG occurs as thin veins and dykes intrusive into anorthosite (Joshi *et al* 2006). The paucity of tonalities or granodiorites in the EGMB suggests that anorthosite massifs and associated alaskitic granites in the EGMB may be the comagmatic product of extensive fractionation of tonalite-granodiorite or equivalent melt (Bose 1979). Sarkar *et al* (1981) demonstrate a gradational chemistry with marked Fe enrichment in the anorthosite–leuconorite–norite–jotunite association (the anorthosite suite), and an abrupt compositional break between the anorthosite suite and the spatially associated high-K acidic suite in the Chilka Lake complex. The authors, by implication consider the anorthositic suite of rocks to be comagmatic and acidic suite to be cogenetic, although they concede extensive hybridisation of the late-stage granitic residual melts of anorthosite crystallization by products of crustal anatexis. In Chilka Lake complex, the initial $^{87}Sr/^{86}Sr$ ratio (0.7066) of anorthosite is quite comparable to that of acidic suite of rocks (table 1). In contrast, Bhattacharya *et al* (1998) interpreted HFSE- and REE-enriched ferrodiorite of Bolangir complex as the late stage melt evolved through crystallization of anorthositic rocks from HAG. On the basis of field relation, petrographic characters and widely varied $\delta^{18}O$ (7–9‰) values, they explain hybrid rocks (monzodiorite) at the border as the mixing of ferrodiorite melt and felsic magma. Thus, the two suites of rocks in the Bolangir complex are not comagmatic (Bhattacharya *et al* 1998). Maji and Sarkar (2004), on the basis of principal component analysis and $\delta^{18}O$‰ values, also proposed a noncomagmatic relation between the two suites of rocks in Turkel complex. Considering major and minor element mass balance computations, they suggested HAG as parental melt, and ferrodiorites to represent the residual melt after crystallization of anorthosite and leuconorites in Turkel complex. The Jugsaipatna massif is unmetamorphosed. However, the other massifs have been metamorphosed in granulite facies conditions. The Chilka Lake anorthosite intruded into a thickened crust at about 28 km depth causing an ultra-high temperature contact metamorphism (Raith *et al* 2007). The high grade supracrustal rocks record at least three phases of deformation. However, the massifs are of different age and set in different regions with different style and age of deformation.

2.2 *North American massif anorthosite complexes*

The eastern shield of North America covering parts of Canada and USA hosts at least 42 anorthosite massifs ranging in age from 1130 Ma to 1650 Ma (Ashwal 1993). Out of five major provinces in this part of the shield, the Greenville province is known to be the largest repository of the anorthosite-mangerite suite of rocks (figure 1c). Most of the massifs except Nain complex, Labrador are metamorphosed to granulite-to-amphibolite grade.

The heart-shaped Marcy massif ($3000\,km^2$) occurring at the core of a broad domal structure is the largest anorthosite body in the Adirondack Mountains and known as 'typical' of Middle Proterozoic massifs worldwide (Ashwal 1993, cf. McLelland *et al* 2004). The massif is enclosed on all sides by a suite of granitoids including mangerite, charnockite, farsundite, leptynite. The core of the massif is dominated by anorthosite, and the other associated rocks are leuconorite, leucogabbro, gabbroic anorthosite and anorthositic gabbro. Towards the border the Fe-Ti oxide-bearing rocks occur. The massif has undergone metamorphism at granulite facies condition (800°C and 9 kbar) and still there is excellent preservation of igneous textures. The anorthositic rocks (ca. 1155 Ma) and the associated granitods (ca. 1158 Ma) in the Marcy massif are coeval (McLelland *et al* 2004). However, the margins of the massif contain a very complex hybrid rocks (Keene gneiss), in which the mineral disequilibria represent commingling of anorthositic and granitoid magmas. The disequilibrium is recorded by the absence of correlation between Mg# [molarMgO/(MgO + FeO)] of pyroxene and An content of plagioclase within Keene gneiss and consistent correlation between the parameters within the anorthosites and granitoids (Buddington 1969; Davis 1969). In addition, the development of reaction rims of perthite on plagioclase xenocrysts incorporated into granitoids and plagioclase over alkali feldspar incorporated into anorthosite experimentally prove the melt-solid disequilibria (Wark and Stimac 1992; Hamilton *et al* 2004).

The Morin massif ($2500\,km^2$) located northwest of Montreal comprises mainly anorthosite and gabbroic anorthosite with minor amount of anorthositic gabbro and ferrogabbro. Based on colour and texture, the anorthositic rocks in this massif is divided into two types, namely, coarse-to-medium grained, mauve or dark colour massive 'mauve facies anorthosite' and 'chertsey facies anorthosite' characterized by fine-to-medium grain size, granulitic-to-gneissic and light green to white colour. Mineralogically these two rocks are similar and they contain variable proportion of unzoned plagioclase, orthopyroxene (hypersthene to bronzite), clinopyroxene (titaniferous augite), interstitial K-feldspar, magnetite, ilmenite and traces of quartz and apatite. At few places within the massif, considerable proportion of subhedral to euhedral hornblende and anhedral red garnet are present. Ferrogabbro occurs as numerous thin dykes and lenses, which cut across the mauve facies rock only. In ferrogabbro, the proportion of oxide, apatite, quartz, K-feldspar and Fe-rich pyroxene is higher than anorthosite, gabbroic anorthosite and anorthositic gabbro. The ferrogabbro shows a protoclastic texture. The mafic minerals are arranged in subparallel manner giving rise to the appearance of flow line. The bordering felsic rocks are mainly mangerite-syenite containing microperthitic orthoclase, plagioclase, orthopyroxene, clinopyroxene, quartz, magnetite and ilmenite.

The Nain massif in Nain province consists of about 7000 km^2 of anorthosite pluton intruded by subequal amount of granitoids. The anorthositic rocks include anorthosite, leucogabbro, leuconorite, troctolite and very minor gabbro and norite. Dioritic rocks often occur as veins/dykes within or adjacent to leuconorite. A hybrid rock representing mixture of granitic and dioritic magma occurs between leuconorite and granite. The massif is typically undeformed and unmetamorphosed except along some faults and fracture zones. Orthopyroxene is the dominant mafic with minor amount of olivine and augite. The Fe-rich dioritic rocks include diorite and jotunite. The granitoid rocks include opdalite, farsundite, adamellite and charnockite. The model melt resembling the Nain anorthosite pluton is typical adakite or Archaean tonalite derived from depleted lower crust of garnet granulite mineralogy and the associated jotunite is not consanguineous (Bedard 2001).

2.3 *Norwegian anorthosite complexes*

The Hidra massif located in the eastern part of the Gerund province, Rogaland, SW Norway, is one of the five massif-type anorthosite rock bodies, which occur in association with a layered lopolith and three large acidic plutons ranging in composition from granite, granodiorite to charnockite (Duchesne *et al* 1985, figure 1d). The major part of the undeformed Hidra massif consists of medium grained leuconorite containing mainly plagioclase, orthopyroxene, Fe-Ti oxide and apatite showing subophitic texture (Demaiffe and Hertogen 1981; Duchesne *et al* 1985). The plagioclase grains are as large as 1 m towards the centre of the massif, the leuconorite grades into coarse grained anorthosite (*sensu stricto*) with well developed orthocumulate texture and interstitial quartz and K-feldspar grains. The leucotroctolite with mainly plagioclase and olivine occurs locally. Pyroxenite rarely occurs as inclusions within leuconorite. The undeformed and unmetamorphosed mineralogy and texture of Hidra massif suggest that it has been emplaced after the major tectonic events that affected the gneissic envelope. A fine grained porphyritic jotunite containing mainly blue plagioclase phenocrysts, hypersthene and Fe-Ti and P-rich accessory minerals occur near the contact of Hidra massif with the adjacent K-rich felsic gneisses. On the basis of field observations, Demaiffe and Hertogen (1981) considered jotunite to be equivalent to the chilled margin of shallower intrusives. A stockwork of charnockite dyke and lensoidal pagmatite cut across the massif. Initial $^{87}Sr/^{86}Sr$ ratio in anorthosite and norite-leuconorite rocks is low (0.7055–0.7057). The jotunites have initial $^{87}Sr/^{86}Sr$ ratios (0.7061), similar to that of plagioclase cumulates (0.7053) as well as anorthosite and norite-leuconorite rocks, implying that they belong to the same magmatic suite (Demaiffe and Hertogen 1981; Demaiffe *et al* 1986). The occurrence of jotunite as fine grained border facies and absence of Eu anomaly suggest that the sequence of rocks in the Hidra massif resulted most likely from a jotunitic parental magma (Duchesne *et al* 1974; Demaiffe and Hertogen 1981; Weis and Demaiffe 1983). On the basis of negative Eu anomaly complementary to the positive Eu anomaly of anorthositic rock and high REE content, they considered the charnockitic rocks to be the residual product from jotunitic melts. The higher initial $^{87}Sr/^{86}Sr$ ratio (0.7086) in charnockite points to contamination by surrounding gneissic rocks.

3. Geochemistry

3.1 *General considerations*

A general survey of the massif anorthosite complexes of India, North America and Norway reveals that mainly three distinct varieties of rocks occur within the massifs, which are bordered by a suite of K-rich silicic-intermediate rocks. For the sake of simplicity in comparing geochemical characteristics under the present study, the rocks within the massifs are referred to as:

- anorthosite (*sensu stricto*) containing >90% plagioclase, after Streckeisen (1976),
- mafic-leucomafic rocks, which include the associated rocks containing <90% plagioclase and described by the earlier workers as leuconorite, leucogabbro, leucotroctolite, gabbroic anorthosite, anorthositic gabbro, etc., and,
- Fe-rich dioritic rocks for fine to medium-grained Fe-rich mafic rocks containing higher proportion of apatite and oxide phases and variably termed earlier as diorites, ferrodiorites, jotunites, ferrogabbro, Fe-rich gabbronorite, ferromonzodiorite and monzonorite. However, the acid to

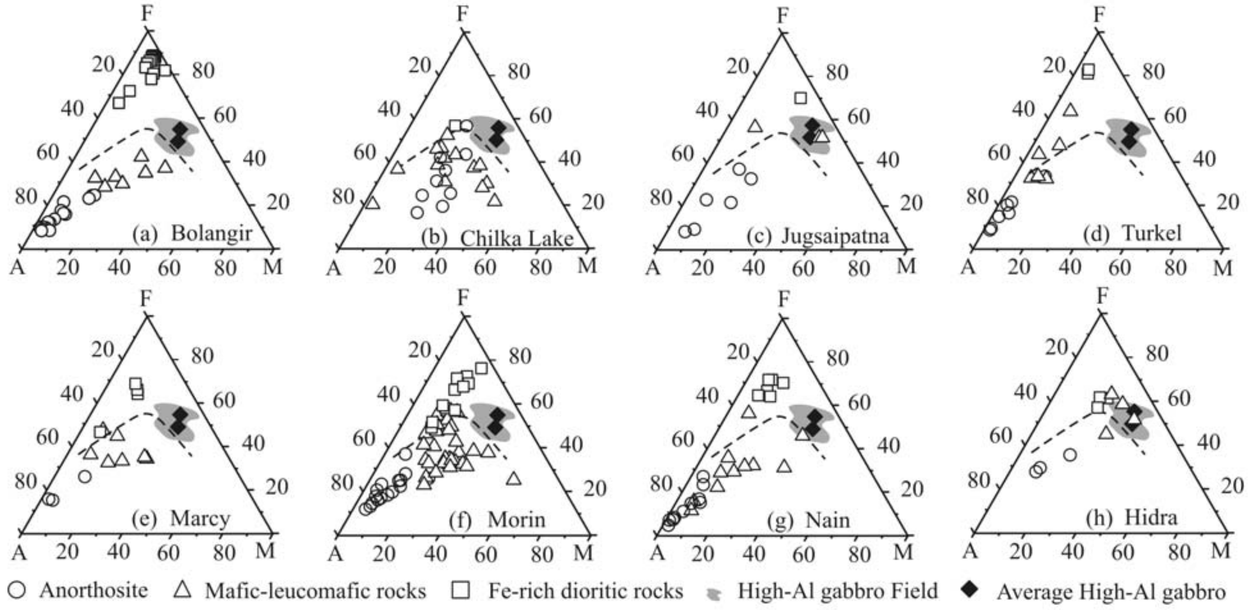

Figure 2. AFM diagram showing variation of whole rock composition of anorthosite, mafic-leucomafic rocks and Fe-rich dioritic rocks for massif anorthosite complexes. The divider suggested by Irvine and Baragar (1971) to separate tholeiitic (above) from calc alkalic (below) suites is shown as dashed line for reference only.

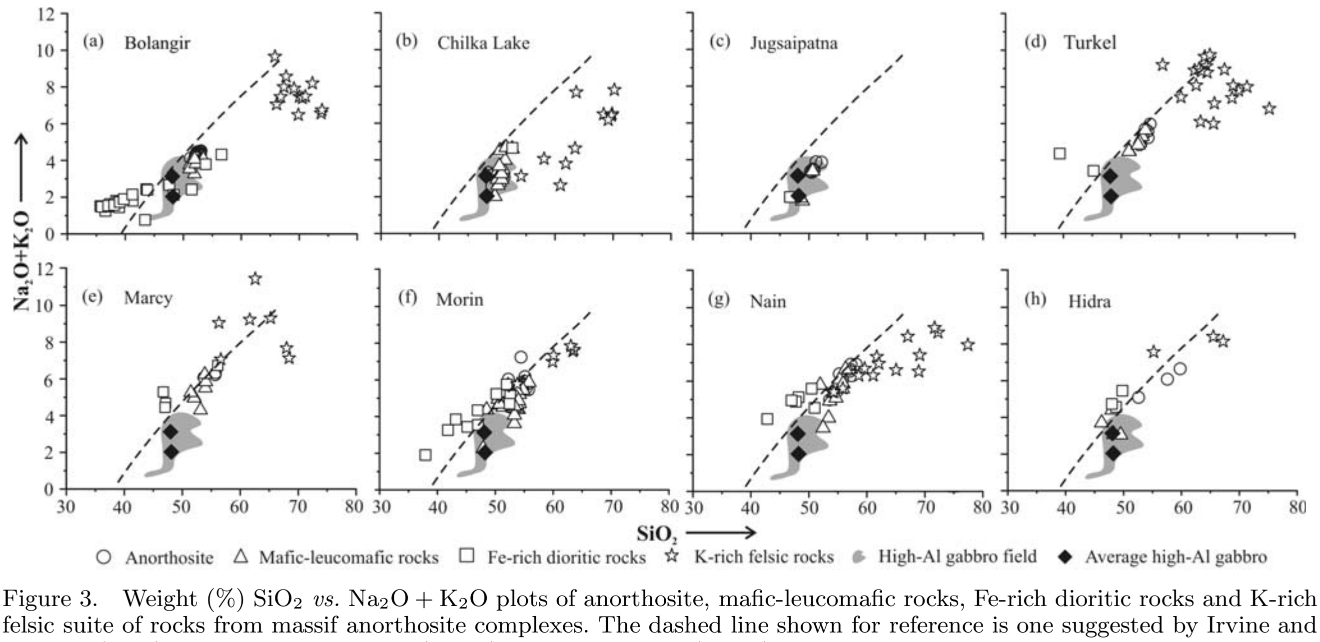

Figure 3. Weight (%) SiO_2 *vs.* $Na_2O + K_2O$ plots of anorthosite, mafic-leucomafic rocks, Fe-rich dioritic rocks and K-rich felsic suite of rocks from massif anorthosite complexes. The dashed line shown for reference is one suggested by Irvine and Baragar (1971) for separating alkaline (above) from subalkaline (below) rocks.

intermediate rocks bordering the massifs such as, granite, granodiorite, monzonite, mangerite, opdalite, farsundite, charnockite and syenite are here termed K-rich felsic rocks.

3.2 *Selection of chemical parameters and variation diagrams*

For a comprehensive appraisal of the geochemical characteristics, the analyses from different published literatures as mentioned earlier are plotted in several bivariate and triangular diagrams (figures 2–10). The REE values cited in the text and shown in the figures are normalized to chondrite values after Taylor and McLennan (1985). It is widely recognized from the mineralogy of the anorthositic rocks that they are characterized by low SiO_2 content and high Al_2O_3 and $Na_2O + K_2O$ compared to FeO and MgO. The range of SiO_2 within the rocks is also narrow. However, it has been found that FeO, MnO, MgO, and TiO_2 show broader variation and strong negative correlation

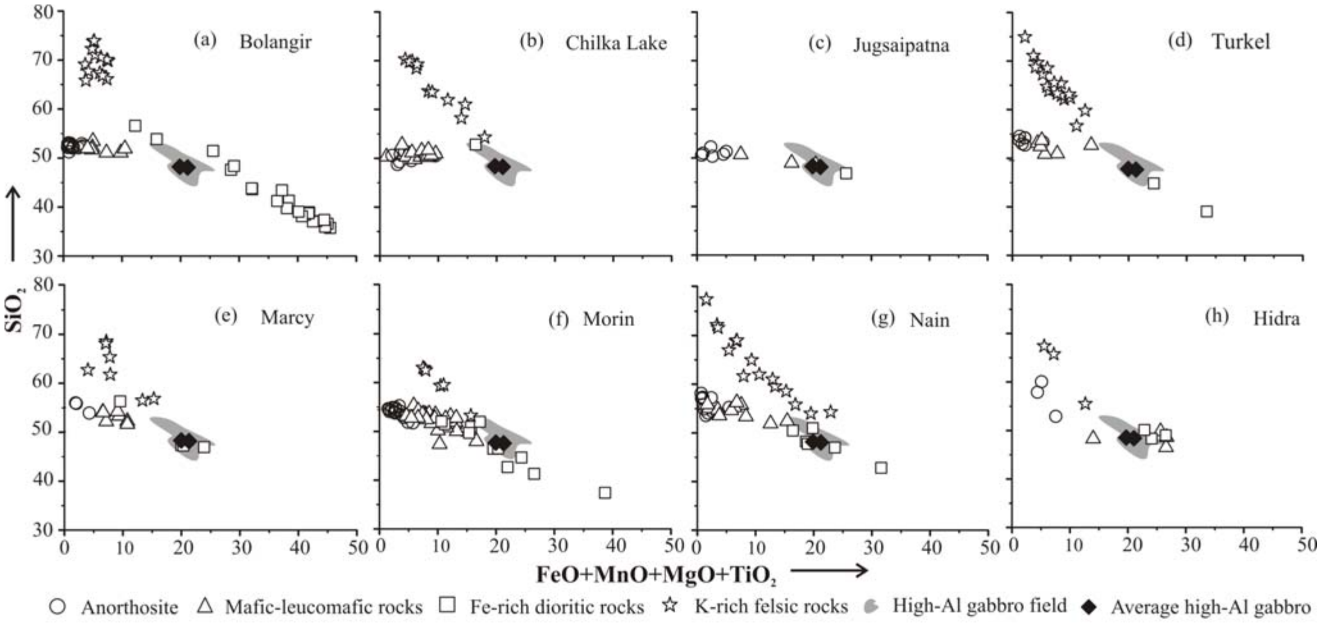

Figure 4. Weight (%) $FeO^{Total} + MnO + MgO + TiO_2$ *vs.* SiO_2 of anorthosite, mafic-leucomafic rocks, Fe-rich dioritic rocks and K-rich felsic suite of rocks from massif anorthosite complexes.

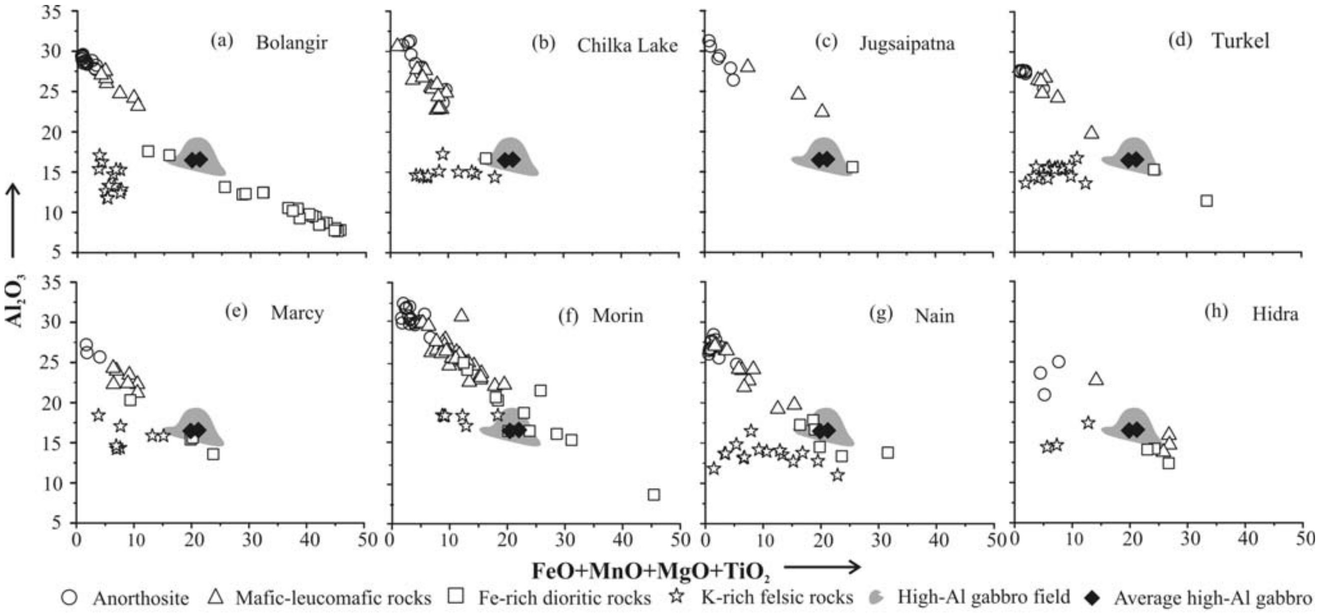

Figure 5. Weight (%) $FeO^{Total} + MnO + MgO + TiO_2$ *vs.* Al_2O_3 of anorthosite, mafic-leucomafic rocks, Fe-rich dioritic rocks and K-rich felsic suite of rocks from massif anorthosite complexes.

with the SiO_2 (figure 4). Hence, in addition to Harker variation plots, a few bivariate diagrams are given with respect to $FeO + MnO + MgO + TiO_2$ to work out the exact nature of chemical variations among the rock types. As anorthositic rocks are supposed to be of crystal cumulate origin and do not represent liquid line of descent, the AFM diagram and the diagram Al_2O_3 wt% *vs.* An/(An + Ab) in plagioclase (Irvine and Baragar 1971) have been used only to indicate relative alkali and mafic oxide enrichment among different members instead of tracing magmatic affinities and fractionation trends like distinguishing calc alkalic or tholeiitic magmas (Ashwal 1993). Normative mineralogy of the rocks were calculated converting total Fe as FeO following the program written by Kurt Hollocher, Department of Geology, Union College, Schenectady, N.Y. 12308, hollochk@union.edu. The An-Ab-Or diagram is to evaluate the composition of normative feldspar.

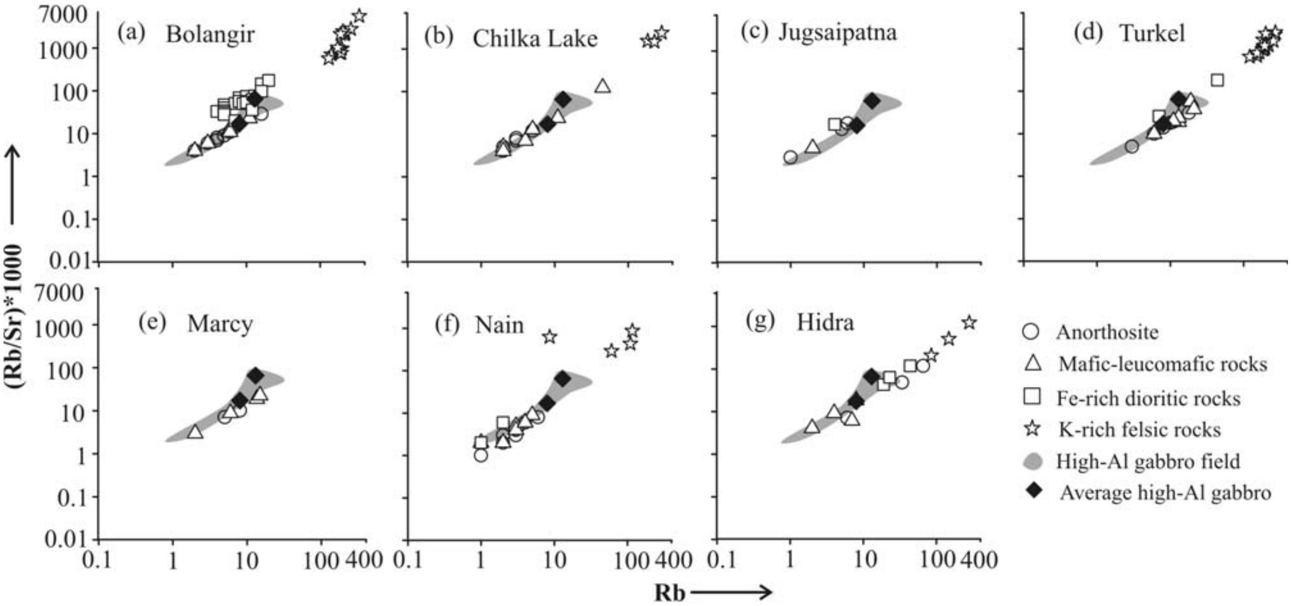

Figure 6. Rb (ppm) *vs.* Rb/Sr × 1000 of anorthosite, mafic-leucomafic rocks, Fe-rich dioritic rocks and K-rich felsic suite of rocks from massif anorthosite complexes.

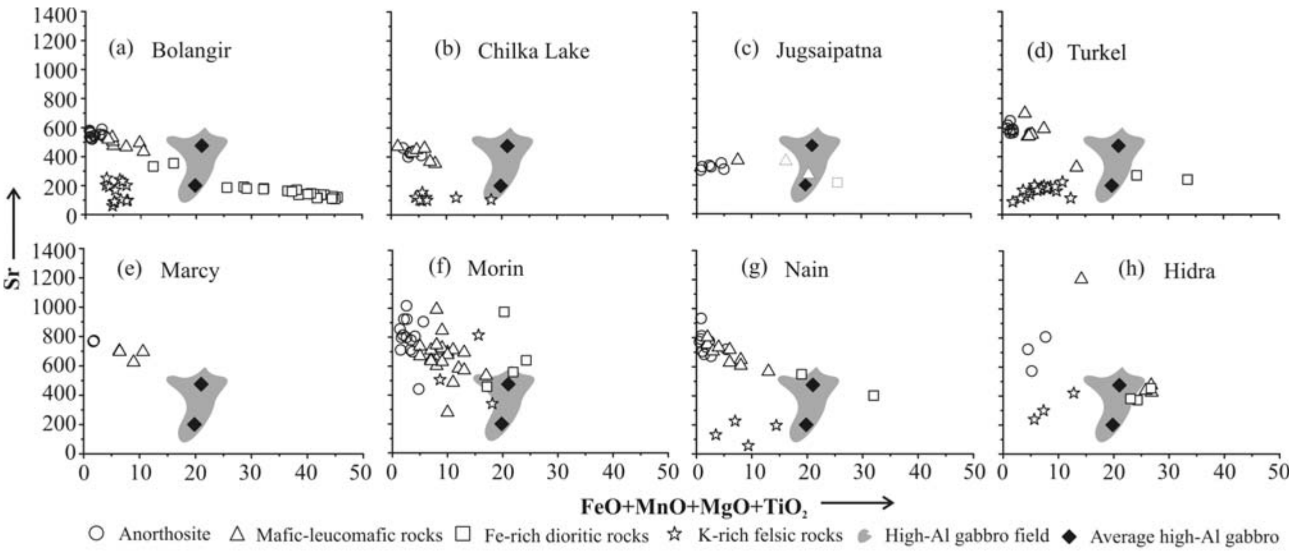

Figure 7. Weight (%) $FeO^{Total} + MnO + MgO + TiO_2$ *vs.* Sr (ppm) of anorthosite, mafic-leucomafic rocks, Fe-rich dioritic rocks and K-rich felsic suite of rocks from massif anorthosite complexes.

Isotopic chemistry ($\delta^{18}O$, initial $^{87}Sr/^{86}Sr$ ratio and ε_{Nd}) and mineral chemical data so far available (table 1) have been used for petrogenetic interpretations.

In several Proterozoic massif anorthosite complexes, the high-Al gabbro (HAG) occurring as chilled margin and thin dykes intrusive in to anorthosite plutons are considered as relics of likely parental melts of anorthosite suite (Mitchell *et al* 1995, 1996). Thus, in variation diagrams, the anorthosite suite of rocks should show a broad spatial relationship with the HAGs. Hence, the compositions of 49 HAG samples reported from Greaser Intrusions, Poe Mountain anorthosites and Strong Creek complex of USA (Mitchell *et al* 1995), and Bolangir, Chilka and Jugsaipatna of India (Joshi *et al* 2006) are plotted in the variation diagrams to judge the suitability of HAG as possible parental magma.

3.3 *Anorthosites*

These rocks are characterized by high Al_2O_3 (20.90–31.29 wt%), $Na_2O + K_2O$ (2.57–7.23 wt%), Sr (304–1017 ppm), and Ba (40–680) contents and low in K_2O (0.13 to 2.34 wt%), FeO^{Total}

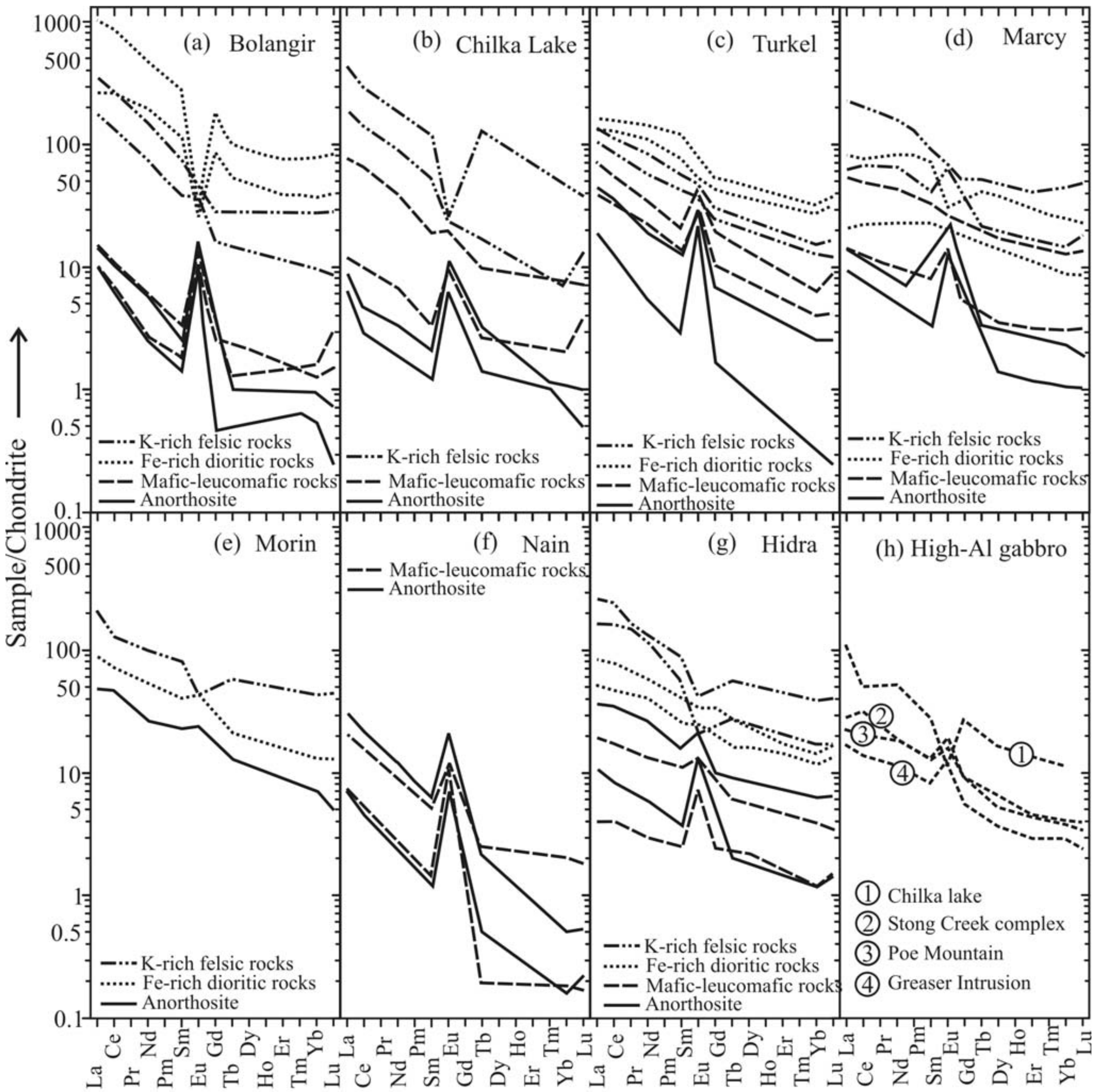

Figure 8. Chondrite normalized (after Taylor and McLennan 1985) REE concentration range in anorthosite, mafic-leucomafic rocks, Fe-rich dioritic rocks and K-rich felsic suite of rocks from massif anorthosite complexes. Average REE character of HAG is also shown for reference.

(0.36–6.61 wt%), TiO_2 (0.07–2.52 wt%), P_2O_5 (0.01–0.35 wt%), V (1–80 ppm), Ni (1–196 ppm), Zn (2–89), Zr (2–84 ppm) contents and Rb/Sr ratio (0.003–0.031). The SiO_2 (48.55 to 59.80 wt%) content is low to moderate and Mg# varies from 16.93–73.79. Normative An/(An + Ab) is within the range of 43.81 to 78.06. Though anorthosites are subalkalic (figure 3), a perusal of the AFM diagram (figure 2) show that they are rich in alkali and Al_2O_3 compared to FeO and MgO. The REE patterns of the rocks are typical of massif-type anorthosite being characterized by LREE enrichment (≈7X–213X chondrite) with very strong positive Eu anomaly (figure 8). In every complex, among all the lithounits, anorthosites have the least REE content. Chondrite normalized La/Lu ratio varies from 3.59 to 126.97.

3.4 *Mafic-leucomafic rocks*

These are subalkalic and characterized by $Al_2O_3 =$ 13.80–30.64 wt%, $Na_2O + K_2O = 1.77$–6.65 wt%,

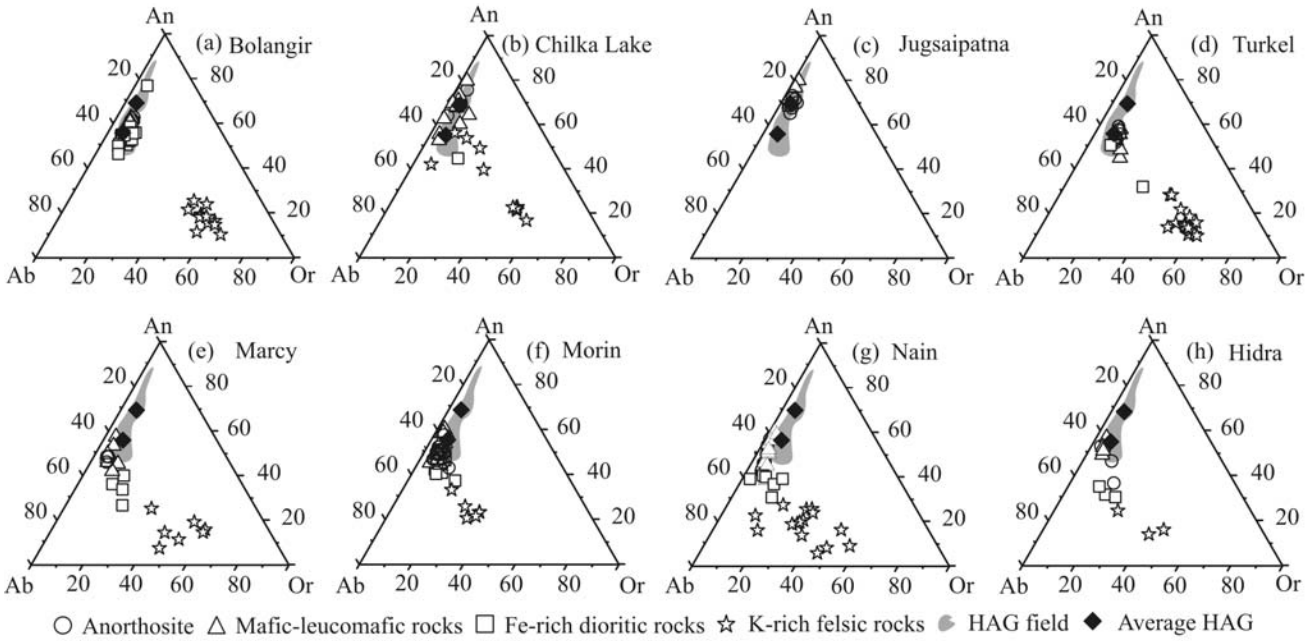

Figure 9. Normative feldspar composition in anorthosite, mafic-leucomafic rocks, Fe-rich dioritic rocks and K-rich felsic suite of rocks in anorthosite-albite-orthoclase triangle from massif anorthosite complexes.

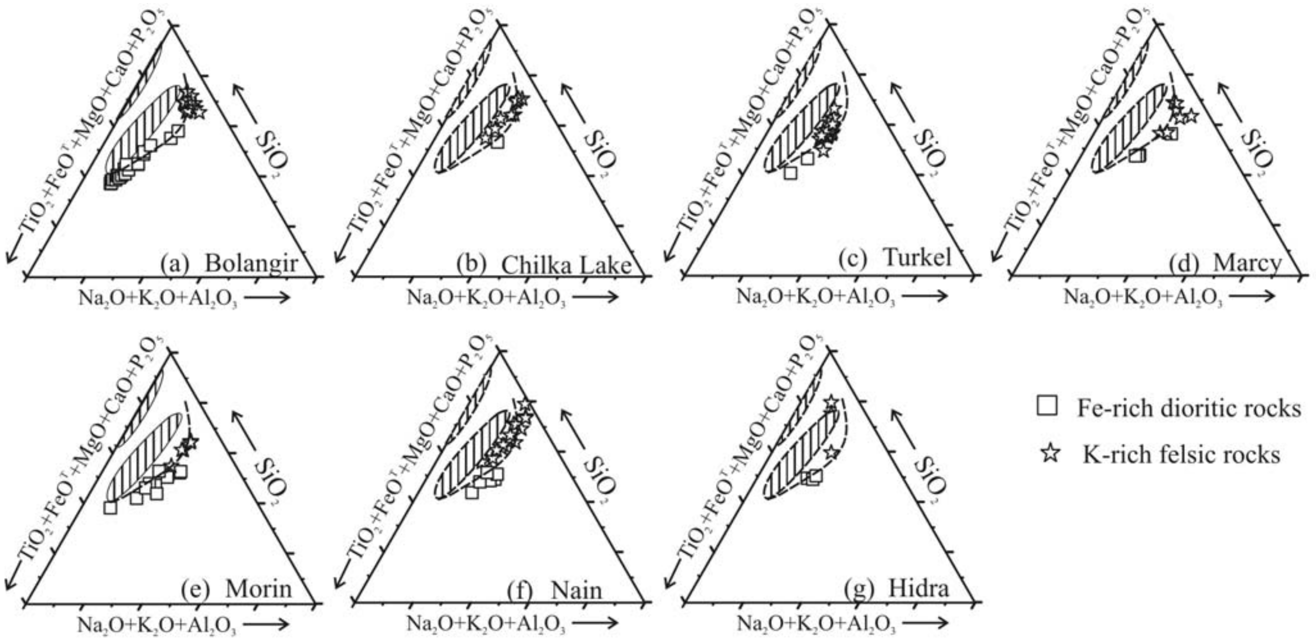

Figure 10. Fe-rich dioritic rock and K-rich felsic rocks from massif anorthosite complexes plotted in pseudoternary diagram after Roedder (1979) showing field for liquid immiscibility in the system leucite-fayalite-silica. The oval-shaped two-liquid field (shaded) is after Weiblen and Roedder (1973) and the liquid immiscibility envelope shown as broken line is from Freestone (1978).

$SiO_2 = 46.20$–56.31 wt%, $K_2O = 0.08$–1.78 wt%, $FeO^{Total} = 0.90$–15.89 wt%, $TiO_2 = 0.09$–5.60 wt%, $P_2O_5 = 0.01$–1.42 wt% and Mg# = 16.00–80.74. Their Sr content (273–1206 ppm) is high. The Ba (25–564 ppm), V (8–425 ppm), Ni (1–183 ppm), Zr (1–928 ppm), and Zn (10–133 ppm) contents are low and variable. Rb/Sr ratio (0.004–0.123) is slightly higher than that of anorthosites. Normative An/(An+Ab) is in between 41.71 and 80.78. The rocks are LREE enriched (up to ≈237X chondrite) with strong positive Eu anomaly (figure 8). Chondrite normalized (La/Lu) ratio varies from 1.96–12.77. The total REE contents of the rocks are comparable to or slightly higher than those of anorthosites. In variation diagrams, the mafic-leucomafic rocks plot in fields either very close or overlapping to the fields of the anorthosites (figures 2–9).

3.5 *Fe-rich dioritic rocks*

In these rocks, SiO_2 ranges from 35.72–56.60 wt% and Al_2O_3 varies between 7.66 and 21.68 wt%. Unlike anorthosite and mafic-leucomafic rocks, these rocks are subalkalic-to-alkalic with very large content of FeO^{Total} (6.94–37.65 wt%), TiO_2 (1.00–10.48 wt%) and P_2O_5 (0.22–2.29 wt%); and very low Mg# (9.00–39.16). $Na_2O + K_2O$ content of the rocks is very low (0.76–6.74 wt%) and they have high Sr (107–975 ppm), Ba (52–1180 ppm), Zr (up to 5512 ppm), Zn (90–489 ppm), V (22–276 ppm) contents. Rb/Sr ratio (0.018–0.182) is higher than that of anorthosites and leucomafic rocks. The normative An/(An+Ab) varies between 35.92 and 81.02 wt%. In ternary feldspar diagram, they have more albite and orthoclase-enriched feldspar compared to the anorthosite and mafic-leucomafic rocks (figure 9). The chondrite normalized REE pattern of Fe-rich dioritic rocks are characterized by high LREE content (up to 1302X chondrite) with strongly negative, zero and negligible Eu anomaly (figure 8). In every complex, these rocks have REE content higher than those of anorthosite and mafic-leucomafic rocks. Chondrite normalized La/Lu ratio varies from 2.33 to 217.28.

3.6 *K-rich felsic suite of rocks*

These rocks are subalkalic-to-alkalic in nature and are mostly granites, although the normative compositions vary up to alkali feldspar granite, quartz monzonite, quartz syenite and syenite (figure 3). They are characterized by high SiO_2 (53.79–77.43 wt%), $Na_2O + K_2O$ (2.62–11.44 wt%), K_2O (0.61–7.59 wt%), Al_2O_3 (11.02–18.41 wt%), Sr (60–420 ppm), Ba (123–2920 ppm) contents; and low in FeO^{Total} (1.28–18.18 wt%) and TiO_2 (0.15–2.42 wt%) contents. Zr content varies between 71 and 1915 ppm. Zn (36–175 ppm) and V (7–180 ppm) contents are moderate. Mg# varies between 5.00 and 56.91. Rb/Sr ratio varies between 0.572 and 5.633. TiO_2 and FeO^{Total} content decrease with the increase of SiO_2 content. Normative feldspar is poor in anorthite and strongly enriched with orthoclase (figure 9). The rocks are LREE enriched (up to ≈537X chondrite) with moderate to richest total REE content in all the complexes. The REE pattern shows Eu anomalies of various sizes and nature (figure 8).

4. Isotopic chemistry

In isotopic chemistry, crustal contamination and resetting of the isotope systems by younger events are evident in most of the massif complexes (Demaiffe and Javoy 1980; Ashwal and Wooden 1983; Emslie *et al* 1994; Morrison and Valley 1998). However, the existing isotopic characters do not deviate much from their original characters (table 1).

Taylor (1969) demonstrated that most 'normal', unmetamorphosed anorthosites have $\delta^{18}O$ values ranging between 5.8 and 7.6‰. Basalts, gabbros and other mantle derived rocks have $\delta^{18}O$ values, between 5.4 and 7.4‰ (Taylor 1968). The $\delta^{18}O$‰ values for anorthosite–mafic-leucomafic rocks reported from Bolangir (6.7–7.5), Turkel (6.9–7.3) and Hidra (5.7–6.4) massifs are well in the range of normal anorthosites and mantle derived rocks (table 1). However, in Hidra massif small departures (5.2‰) from the 'normal range', call for some kind of crustal contamination, with ^{18}O depleted roof rocks (Demaiffe and Javoy 1980). Also in Marcy massif, the $\delta^{18}O$ values are highly variable, enriched up to 9.5‰ through crustal contamination and depleted to 3.0‰ due to isotopc exchange with heated meteoric water when the anorthosite intruded at shallow depth (Morrison and Valley 1998). The $\delta^{18}O$‰ for felsic rocks in Bolangir (7.5–9.2), Turkel (8.0–10.8) and Hidra (6.8–7.4) are higher than their corresponding anorthosite–mafic-leucomafic rocks suggesting either crustal contamination or a crustal origin for these rocks (Demaiffe and Javoy 1980; Bhattacharya *et al* 1998; Maji and Sarkar 2004). The $\delta^{18}O$‰ ranges for Fe-rich dioritic rocks in these three complexes are very close to or slightly higher than anorthosite–mafic-leucomafic rocks (table 1).

Heath and Fairbairn (1969) found that the initial $^{87}Sr/^{86}Sr$ in anorthositic rocks varies from 0.703–0.706, comparable to the mantle derived basaltic rocks. In the present study, the initial $^{87}Sr/^{86}Sr$ in anorthosite–mafic-leucomafic rocks ranges between 0.7034 and 0.7066, close to the range observed by Heath and Fairbairn (1969). The felsic rocks have higher initial $^{87}Sr/^{86}Sr$ ratio (0.7067–0.7104), indicating crustal contamination or derived from the crustal source rocks. In Fe-rich dioritic rocks, the initial $^{87}Sr/^{86}Sr$ ratio ranges from 0.7042–0.7061, close to that of anorthosite–mafic-leucomafic rocks.

The anorthosite–mafic-leucomafic and Fe-rich dioritic rocks in Marcy and Hidra complexes have ε_{Nd} values between +0.44 and +5.5 implying depleted mantle source for these rocks. However, in Nain complex, the negative ε_{Nd} values for anorthosite–mafic-leucomafic rocks, Fe-rich dioritic and K-rich felsic rocks extend over broadly similar ranges (table 1). This indicates that the relative proportion of enriched mantle and crustal Nd isotopes display similar magnitude in all three compositional varieties (Emslie *et al* 1994). In ε_{Nd} *vs.* initial $^{87}Sr/^{86}Sr$ plot, Fe-rich dioritic rocks tend to occupy similar fields to the anorthosite–mafic-leucomafic rocks while the field for K-rich felsic suite though slightly overlapped, extends far away from the field of other two. Thus, mixing between mantle and crustal melt might explain all of the observed variations (Emslie *et al* 1994). Mingling between granitoid and ferrodiorite melts has been documented in the southern part of Nain complex (Wiebe 1978).

5. Petrogenetic implications

The wide composition variation in the whole rock chemistry of anorthosite, mafic-leucomafic rock and Fe-rich dioritic rocks in individual massif as well as among the massifs is due to the large variation in the composition of plagioclase cumulates and mafic minerals (table 1) suggesting crystallization of the minerals for a protracted period over a broad P–T range. The variation may also be attributable to the variable contamination with crustal materials and/or fluids, post-magmatic fluid, and metamorphism. However, the data scatter is not that much to obscure the exact geochemical signatures to make the plots non-conclusive. The effect of metamorphism in major and trace element variation is negligible as the variations in the metamorphosed and unmetamorphosed massifs are quite similar. Because anorthosites, mafic-leucomafic and Fe-rich dioritic rocks contain plagioclase cumulates, the frequent overlapping values of certain geochemical parameters in these rocks suggest fractionation of cumulates to a limited extent.

A close scrutiny of whole rock chemical data and element concentration plots shown in figures 2–10 indicates that within the massifs, FeO, P_2O_5, TiO_2, $FeO^{Total} + MnO + MgO + TiO_2$, Ba, V, Zr, Rb/Sr and REE contents gradually increase while $Na_2O + K_2O$, Al_2O_3, SiO_2, Mg# and Sr contents systematically decrease from anorthosite to Fe-rich dioritic rocks through mafic-leucomafic rocks. The anorthosite, mafic-leucomafic and Fe-rich dioritic rocks are not K-rich. However, their K content increases with decreasing anorthite content of plagioclase (figure 9). On the contrary, subalkalic-to-alkalic K-rich felsic rocks are characterized by high SiO_2, $Na_2O + K_2O$, K_2O, Al_2O_3, Sr, Ba, Zr contents and Rb/Sr ratio; and low in FeO^{Total}, TiO_2 and Mg#. All the rock units in the complexes are LREE enriched (figure 8). The anorthosites are characterized by pronounced positive Eu anomalies. The mafic-leucomafic rocks have REE abundances comparable to or slightly higher than anorthosites with negligible to pronounced positive Eu anomalies. The Fe-rich dioritic rocks have no or negative Eu anomalies. The REE patterns of K-rich felsic rocks fall into two groups. They have REE content either the richest among all the rock suites or in between the anorthosite–mafic-leucomafic rocks and Fe-rich dioritic rocks. The nature of the Eu anomaly in the K-rich felsic rock is also highly variable. Chondrite normalized La/Lu ratio increases from anorthosite to Fe-rich dioritic rocks. Anorthosite, mafic-leucomafic and Fe-rich dioritic rocks plot either in close or overlapping fields defining almost linear trends (figures 3–7). The K-rich felsic suite of rocks falls off the trends.

The anorthosite and mafic-leucomafic rocks contain variable proportion of plagioclase, orthopyroxene, clinopyroxene, olivine, ilmenite, magnetite, apatite, biotite, hornblende, quartz and/or coronal garnet showing typical cumulus texture. The fine-grained and often porphyritic texture-bearing Fe-rich dioritic rocks also have the similar mineralogy but there is a systematic increase in mafic silicates and oxide minerals content, decrease in plagioclase content, An content of plagioclase and X_{Mg} of orthopyroxene from anorthosite to Fe-rich dioritic rocks. In contrast, the K-rich felsic rocks contain perthitic K-feldspar>antiperthitic plagioclase, quartz, orthopyroxene, clinopyroxene, pre- to syn-kinematic garnet, ilmenite, apatite, zircon, rutile, hornblende and biotite showing commonly rapakivi texture. The garnet is often porphyroblastic containing numerous rounded inclusions of quartz, plagioclase, K-feldspar, biotite and ilmenite.

The $\delta^{18}O$ values (5.7–7.5‰), in agreement with the initial $^{87}Sr/^{86}Sr$ (0.7034–0.7066) and ε_{Nd} values (+1.14 to +5.5) suggest a depleted mantle origin for anorthosite, mafic-leucomafic rocks. However, the K-rich felsic rocks having higher $\delta^{18}O$ values (6.8–10.8‰) and $^{87}Sr/^{86}Sr$ (0.7067–0.7104) represent involvement of crustal rocks or crustally derived melt. The Fe-rich dioritic rocks have isotopic characters very close to that of anorthosite–mafic-leucomafic rocks. In ε_{Nd} *vs.* initial $^{87}Sr/^{86}Sr$ plot, Fe-rich dioritic rocks tend to occupy similar fields to the anorthosite–mafic-leucomafic rocks while the field for K-rich felsic suite though slightly overlapped, extends far away from the field of other two (Emslie *et al* 1994).

Thus, the above observations strongly suggest that the anorthosite–mafic-leucomafic rocks are mineralogically and chemically distinct from the K-rich felsic rocks and are not consanguineous. It stands to reason, therefore, that the comagmatic nature of the low-K anorthosite–leuconorite suite and the high-K quartz monzonite and granitic suite of rocks advocated for the Chilka Lake by Sarkar *et al* (1981) is not tenable. It also follows that the argument made by Bose (1979) in a review that the low-K and high-K suites in the Eastern Ghats massif anorthosite complexes share a common parentage is also questionable.

6. Magmatic lineage of ferrodiorite and parental magma

The genesis of Fe-rich dioritic rocks in relation to the genesis of massif anorthosites and the associated K-rich felsic rocks can be viewed as either they are residual liquids after extensive crystallization of plagioclase and other minerals from mantle-derived basaltic magmas or represent segregated Fe-rich melts formed by liquid immiscibility of Fe-rich (ferrodiorite?) and Si-rich (quartz monzonite?) liquids without any genetic link to the anorthosite–leucomafic suite. In addition, monzodioritic or jotunitic melts (equivalent to the Fe-rich diorite) are rarely considered to be the parental magma for anorthositic and bordering charnockitic rocks as in Hidra massif.

In most of the variation diagrams (figures 4–7), the fields of Fe-rich dioritic rocks follow both the trends defined by anorthosite–mafic-leucomafic and high-K felsic rocks. Thus, the Fe-rich dioritic rocks seem to belong to either/both of the suites. Hence, to trace the possible magmatic linkage of dioritic rocks, the suitability of either of the above three explanations in literature is tested. The best way to deduce the role of liquid immiscibility in silicate melt is with a pseudoternary diagram of the system leucite–fayalite–silica (Roedder 1979). In course of the present study, the K-rich felsic suite of rocks plot towards SiO_2 end and Fe-rich dioritic rocks plot towards $CaO + TiO_2 + MgO + FeO + P_2O_5$ with lack of compositional gap between the rock suites for all the complexes (figure 10). In addition, the Ba content in ferrodiorites (52–1180 ppm) is lower compared to that of K-rich felsic suite of rocks (123–2920 ppm) unlike the experimental results of Ellison and Hess (1991) on liquid immiscibility. Thus, Fe-rich dioritic rocks and K-rich felsic suite of rocks do not appear to be genetically related via liquid unmixing.

Alternatively, the ferrodiorites are parent melts to anothositic rock or late-stage residual melts that are residual to extensive anorthosite–mafic-leucomafic crystallization. In variation diagrams, the disposition of anorthosite–mafic-leucomafic rocks and Fe-rich dioritic rocks on either side of collinear or nearly collinear trends and K-rich felsic rocks falling off the trends make the Fe-rich dioritic rocks to be unsuitable to represent parent magma for the two suites of rocks. It has also been discussed that the anorthosite–mafic-leucomafic suite of rocks and K-rich felsic suite are not consanguineous. Hence, their genetic linkage to a common parentage of jotunitic melt can be ruled out. Moreover, the intermediate jotunitic parent magma are considered to be the product of melting of mafic rock belonging to the lower crust (Longhi 2005), but the observed $\delta^{18}O$‰, initial $^{87}Sr/^{86}Sr$ and ε_{Nd} value in anorthosite–mafic-leucomafic rocks suggest their derivation from depleted mantle source. Experimental work of Fram and Longhi (1992) also shows that the P–T values (12 kbar and 1200°C) require to generate anorthosite parent melt are unlikely to be realised at lower crustal conditions. Thus, it appears that jotunitic parent magma is not to be realistic.

On the other hand, the main chemical characters of ferrodiorites, e.g., low SiO_2, MgO; moderate-to-low Al_2O_3; moderate-to-fairly high CaO; high FeO^{Total}, TiO_2, P_2O_5, Zr, Rb/Sr and REE content are also characteristics of late stage residual products of strongly fractionated tholeiitic magma in anorogenic settings (Emslie 1978). Further, mineralogical features such as sharp decrease in X_{Mg} of ferromagnesian phases and An content of plagioclase from anorthosite–mafic-leucomafic rocks to Fe-rich dioritic rocks (table 1) would be compatible with the derivation of Fe-rich dioritic melts by continuous crystallization of anorthosite and mafic-leucomafic rocks. The close similarity between anorthosite–mafic-leucomafic rocks and Fe-rich dioritic rocks in their initial $^{87}Sr/^{86}Sr$ and ε_{Nd} value also support their genetic linkage. However, the $\delta^{18}O$ enrichment of ferrodiorites compared to the anorthosite and mafic-leucomafic rocks in Bolangir complex is incompatible with plagioclase fractionation alone, as it would drive the residual ferrodiorite melt to lower $\delta^{18}O$ values. If the observed higher values in ferrodiorites resulted from fractional crystallization then some phases like ilmenite, which preferentially takes in ^{16}O must coprecipitate with plagioclase. The intimate association of Fe–Ti oxides cumulates with ferrodiorites in many maassifs give the evidence of such fractionation process (Mitchell *et al* 1996). Another effective process in increasing $\delta^{18}O$ values in ferrodiorites may be the isotope exchange with the adjacent felsic rock rich in ^{18}O. Thus, the anorthosite–mafic-leucomafic rocks and Fe-rich dioritic rocks constitute a comagmatic suite. By contrast, the K-rich felsic intrusives bordering the massifs being characterized by the peraluminous nature, high contents of SiO_2, K_2O, Ba and Zr, high K_2O/Na_2O and Rb/Sr ratios, low Mg# and poikiloblastic nature of garnets are suggestive of crustal origin presumably derived by high degree of incongruent melting. Their isotopic characters such as, higher $\delta^{18}O$ values and initial $^{87}Sr/^{86}Sr$ ratio also support this sort of genesis. The ill-defined compositional trends shown by K-rich felsic rocks possibly suggest wide variety of metamorphosed sedimentary/volcano-sedimentary piles of precursor rocks with widely varied compositions.

The composition fields of HAG have overlapping relations to the compositions of anorthosites–mafic-leucomafic rocks and Fe-rich dioritic rocks (figures 4–7) suggesting the parental liquid very close to the composition of HAG. Low initial $^{87}Sr/^{86}Sr$ (0.7033) and ε_{Nd} (up to +2) in HAG similar to that of anorthosite–leuconorite–ferrodiorite also support its genetic linkage with those rocks. The $\delta^{18}O$ range (5.2–7.5‰) in anorthosites–mafic-leucomafic rocks is also close to that mantle derived gabbro (5.4 and 7.4‰). High content of oxides and apatites, strong enrichment of Fe, P, and Ti along with depletion of Si in Fe-rich dioritic rocks representing residual melt suggest that anorthositic rocks must have crystallized in reduced condition. The oxygen fugacity for anorthosite, leuconorite (Jugsaipatna) and orthopyroxene cumulates (Nain) ranges between 0.24 log unit above or close to FMQ and 1 log unit below FMQ during cryatallization of anorthositic rocks is much lower than those recorded in crustal rocks indicating the derivation of parental melts in the mantle (Mitchell *et al* 1996). The HAG is of tholeiitic nature characterized by positive and negative Eu anomalies (figure 2). It is completely different from HAG generated in subduction environment (Myers *et al* 1986; Myers 1988). Thus, the HAG related to the Proterozoic anorthosite complexes have an origin other than that of subduction related HAG. Such HAG in Proterozoic anorthosite complexes generates from tholeiitic magma by resorption of plagioclase and later fractionation of pyroxenes (Markl and Forst 1999). The exact bulk liquid representing the anorthositic suite of rocks have originated from HAG after initial fractionation of some Fe–Mg silicates. This process may produce a wide variety of HAG, depending on the amount of fractionation in tholeiitic melt before resorption of plagioclase, and bulk composition of resorbed plagioclase and fractionated pyroxene. However, the HAGs are normally with positive Eu anomaly because of the large scale resorption of plagioclase. The positive Eu anomaly of HAG explains the zero to low Eu anomaly in ferrodiorites though it represents the residual melts.

Thus, the present study opposes the consideration of monzodiorite (=jotunite) to represent parental liquid for Hidra massif as invoked by Demaiffe and Hertogen (1981) based on the absence of Eu anomaly in jotunite. The authors' other arguments like (i) fine grained nature of jotunite, (ii) $^{87}Sr/^{86}Sr$ ratios (0.7055) similar to the plagioclase cumulates, and (iii) the absence of Eu anomaly in jotunite, and (iii) large negative Eu anomaly in charnockite contemporary to the positive Eu anomaly in anorthosite and leuconorite to represent the jotunite as parent melt are not also unequivocal. The jotunite could even show the same characters if it would have been the residual melts derived from a parent melt/HAG with positive Eu anomalies. Moreover, enrichment of plagioclase imcompatible elements, increase of REE abundances from anorthosite to ferrodiorites through leuconorite suggest that jotunite of Hidra massif represents a residual melt from crystallization of anorthosite and leuconorite.

7. Plausible magmatic evolution of the massifs and the bordering K-rich felsic suites

Considering the two-stage model of Ashwal (1993) and mass balance computation of Maji and Sarkar (2004), the above arguments based on synthesize of field, petrological and chemical data leads to suggest the following magmatic evolutionary histories for the Proterozoic anorthosite complexes. The mantle derived-HAG melt tends to pond below a continental lid at the crust-mantle boundary with voluminous crystallization of plagioclase in the magma pools. The buoyant plagioclase-rich crystal mush during ascent trap pools of resident magmas. Decompression in the trapped magma pools led to more plagioclase crystallization, forcing the residual melt to crystallize pyroxene + plagioclase to produce the leucomafic rocks and then continued crystallization of plagioclase + pyroxene. At the end stage of this process the residual melt became highly enriched in Fe and plagioclase-incompatible elements. Mineralogical and geochemical characters indicate that in most cases, either apatite and zircon saturation limits in the melts were eventually reached or these minerals were assimilated from the surrounding crust (Maji and Sarkar 2004). The dense Fe-rich dioritic melts supported by the densely-packed plagioclase crystal mush were subsequently extracted to the low-pressure repositories at or near the roof of the ascending anorthosite diapir due to tectonic forces acting on the near-solidified crystal mush. However, the extraction of residual melts from their hosts was not effective enough. This resulted in Fe-rich melts to inherit early formed crystals from the coexisting mafic-leucomafic rocks and thereby giving rise to an overlapping chemistry with the anorthosite and mafic-leucomafic rocks.

The K-rich felsic suites bordering the massifs are generated by incongruent melting reactions of lower crustal rocks. Heat supply for the formation of the crustal melts was possibly provided by the release of latent heat during plagioclase crystallization in the ascending basic magma pools. The wide compositional diversity in the felsic rocks may be attributed to the different types of incongruent melting that occurred in the different precursors. The felsic melts provided lubrication for the anorthosite–mafic-leucomafic diapir rising in their wake *en route* to the final emplacement of the igneous intrusives.

8. Conclusions

In Proterozoic anorthosite complexes, similar rock types have widely varied but similar mineralogy, and major and trace element composition. Within the massifs, anorthosites (*sensu stricto*)–mafic-leucomafic rocks are subalkalic and are characterized by high $Na_2O + K_2O$, Al_2O_3, SiO_2, moderate to low Mg#, low abundances of plagioclase-incompatible elements and REE with strongly positive Eu anomaly. Their $\delta^{18}O$‰, initial $^{87}Sr/^{86}Sr$ ratio and ε_{Nd} indicate a depleted mantle source. By contrast, the K-rich felsic rocks bordering the massifs being characterized by the peraluminous nature, high contents of SiO_2, K_2O, Ba and Zr, high K_2O/Na_2O, Rb/Sr ratios REE, $\delta^{18}O$‰ and initial $^{87}Sr/^{86}Sr$ ratio are suggestive of crustal origin. The subalkalic to alkalic Fe-rich dioritic rocks occurring at the interface between the massifs and bordering K-rich felsic rocks have isotopic, chemical and mineral composition more closely akin to anorthosite–mafic-leucomafic rocks than the felsic rocks. However, there is a gradual decrease of plagioclase content, An content of plagioclase, X_{Mg} of pyroxenes and an increase of mafic silicates and oxides from anorthosite–mafic-leucomafic rocks to Fe-rich dioritic rocks. They are richer in Fe, Mn, Ti, P, plagioclase-incompatible high field strength elements and REEs with negligible/negative Eu anomaly. The Fe-rich dioritic rocks represent residual melt from the magma which produced anorthosite–mafic-leucomafic rocks. They neither represent members of immiscible pair conjugate to bordering felsic rocks nor parent magma. The magmas parental to the anorthosite suite are fractionated mantle-derived melts, resembling high-Al gabbro of varied composition.

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