The Turkel anorthosite complex revisited

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Geological investigation in recent years reveals that the anorthosite-leuconorite massif (81 sq km) is much larger than known from previous studies. The massif is bordered by a suite of garnetiferous felsic rocks comprising quartz monzonite gneiss, granite gneiss and megacrystic K-feldspar-bearing granite. Ferrodiorites, hitherto unknown from this area, occur as veins at the massif-felsic suite interface, and as rare apophyses within leuconorites at the massif margin. The massif and the bordering felsic rocks were presumably emplaced during the earliest of the three phases of folding documented by the metasedimentary gneisses that host the massif.

The petrographic and geochemical characteristics suggest that the low-K anorthosite-leuconorite-ferrodiorite suite does not share a common parentage with the bordering high-K felsic intrusives. The anorthosites and leuconorites were derived by polybaric fractionation of mantle-derived melts. The ferrodiorites are anorthosite residual melts that were not entirely segregated from the host solids. By contrast, the granite gneisses and granites originated by incongruent melting of crustal rocks. The chemical differences between quartz monzonite and granite gneisses point to their derivation from different crustal precursors.

1. Introduction

Anorthosite-mangerite-charnockite-granite suite of rocks are reported from virtually all high grade terranes of the Proterozoic age. The petrologic and geochemical evolution of these enigmatic rock suites and their tectonic setting have been endlessly debated by petrologists and geochemists. The massif-type anorthosites in the Eastern Ghats terrane occur in an arcuate belt that parallels the tectonic contact between the granulite facies gneisses of the Proterozoic age that host the massifs in the east, and the Singhbhum and Bhandara cratons of the Archaean age to the north and west. However, inspite of the known occurrences of the Eastern Ghats anorthosite massifs since the turn of the century (Walker 1902), very little is known about their field setting, the petrologic phase relations and the geochemical characteristics of the rock suites. In the present communication, an attempt has been made to understand the geochemical evolution of one such massif anorthosite complex at Turkel.

2. Field setting

A 630 sq km area around Turkel near the northwestern fringe of the Eastern Ghats Belt is geologically mapped in the scale of 1:50,000 (figure 1). The area comprises a centrally-bulged, NE-SW trending spindle-shaped anorthosite-leuconorite massif. The massif extends over an area of 81 sq km, and is much larger than was proposed by earlier researchers (6.5 sq km, Chatterjee 1965). Garnetiferous quartz monzonite gneisses, granite gneisses and megacrystic K-feldspar-bearing granites border the massif from West through North to East. Ferrodiorite occurs as veins at the interface between the massif and the quartz monzonite gneisses, and as rare apophyses within leuconorites. The ferrodiorites were hitherto

Keywords. Turkel (Orissa); massif anorthosite; field setting; petrography; geochemistry.
unknown from the Eastern Ghats anorthosite massifs. In the southern and the southeastern parts of the area, the massif is cushioned against an assemblage of meta-pelitic/semipelitic gneisses, calc-silicate gneisses, migmatitic quartzofeldspathic gneisses (leptynite) and bands/boudins of mafic granulites. The metasedimentary gneisses also occur as structurally conformable enclaves within the felsic intrusives. A small septum of accordantly folded leptynite-mafic granulite-calc-silicate gneiss composite occurs within the massif.

Three folding events are documented by the metasedimentary gneisses. The earliest deformation ($D_1$) is represented by rootless hinges of intrafolial folds that produced the pervasive gneissic layering in the rocks. The $S_1$ surface was refolded successively by near-coaxial tight to isoclinal reclined $D_2$ folds and open, upright $D_3$ folds with their fold axes plunging at shallow to moderate angles towards east. The development of planar fabric due to $D_2$ folds is not evident, but brittle fractures paralleling the axial planes of $D_3$ folds are observed in the metasedimentary gneisses. Although no mesoscopic folds are observed in the felsic intrusives, the rocks are characterized by strongly developed gneissic foliation, manifested by schleric aggregates of ferromagnesian phases, strongly drawn-out K-feldspar augen, lenticular quartz grains and flattened garnet porphyroblasts. Clearly, the rocks experienced strong deformation post-dating solidification. The $\pi$-pole distribution of gneissic foliation in the felsic intrusives is conformable with the $\pi$-pole girdle of $S_1$ planes in the metasedimentary gneisses. This indicates that the felsic gneisses were emplaced prior to or during $D_1$. The ferrodiorites are structurally concordant with the quartz monzonite gneisses, but are intrusive within the massif. Accordingly, a pre- to syn-$D_1$ emplacement age for the massif is suggested here. In the accompanying paper by Raith et al (1997), the Bolangir massif is deemed to be of pre-$D_2$ emplacement age, although a syn-$D_1$ age is not entirely precluded by the authors. Also, in a recent communication, Dastidar et al (1997) conclude that the anorthosite massif at Bankura predates the two folding events documented by the bordering metasedimentary gneisses. By contrast, the massive to
weakenly foliated K-feldspar-bearing granites at Turkel occur as boudins in the limbs of D₂ folds and as segregation at D₁ fold hinges. Also, they contain randomly-oriented xenoliths of foliated leptynite. Accordingly, a pre- to syn-D₂ emplacement age is assigned to the rocks.

3. Petrography

Anorthosites, leuconorites and ferrodiorites contain the following minerals, plagioclase (An₅₃₋₃₉), inverted pigeonite (Xₐ₉ ≈ 0.52₋₀.2₉), ilmenite, magnetite and apatite with accessory amounts of biotite, hornblende, interstitial K-feldspar and quartz. Within the suite, the modal percentages of ferromagnesian silicates and apatite increase dramatically towards the ferrodiorite; also, Fe/Mg ratio in pyroxenes increases sharply, and An content in plagioclases decreases. Fayalitic olivine (Xₐ₉ ≈ 0.1₁) and zircon appear at the ferrodiorite end. The rocks are characterized by protoclastic to granoblastic textures. However, relic cumulus textures in the anorthosites and leuconorites, and porphyritic textures defined by plagioclase phenocrysts in ferrodiorites are occasionally observed in the less deformed varieties. Garnets (Alₙ₉₋₇₁Prₚ₃₋₁₂Grₛ₂₁₋₂₄ Spₛ₂₋₄) with dactylitic quartz intergrowths typically mantled inverted pigeonites and/or pyroxene clusters. The garnets are clearly of metamorphic origin (3Fs + 3An = 2Alm + Grs + 3Qtz and 3Hd + 3An = Alₙ₉ + 2Grs + 3Qtz) and mostly restricted to the leuconorites and ferrodiorites (mineral abbreviations are after Kretz 1983). The coronal garnets are occasionally decomposed to orthopyroxene (Xₐ₉ ≈ 0.₄₁) and calcic-plagioclase (An₉₅) intergrowths, 2Alm + Grs + 3Qtz = 3Fs + 3An.

Quartz monzonite and granite gneisses are characterized by polygonal granoblastic textures that overprint the characteristic gneissic fabric in the rocks. Minerals present in decreasing order of abundance are scapolite (Ca/Ca+Na = K = 0.₈₉; CO₃/CO₃ + SO₄ + Cl = 0.₉₅), plagioclase (An₉₈), clinopyroxene (Xₐ₉₋₀.₄₈), wollastonite, K-feldspar and sphene. The significant reaction textures observed in the rocks are (i) decomposition of melanoctic scapolite to plagioclase + calcite intergrowths (Scp=3An + Cal), (ii) stabilization of coronal garnet, virtually grossular in composition (Alₙ₉Grₛ₉₆Spₛ₂), with blebs/warts of calcite and quartz at the interface between scapolite and wollastonite/calcite (Scp+6Wo=3Grs+3Qtz+Cal, Scp+5Wo=3Grs+2Qtz+CO₂ and Scp+3Wo+2Cal = 3Grs+3CO₂), and (iii) the formation of coronal grossular at the interface between plagioclase-calcite intergrowths that replace scapolite (An = 2Cal + Qtz = Grs + 2CO₂).

Meta-pelite/semipelite gneisses have a simple, but highly variable mineralogy: quartz-K-feldspar-sillimanite-garnet-ilmenite ± cordierite ± biotite ± spinel ± rutile ± plagioclase ± graphite ± muscovite. The rocks are characterized by polygonal granoblastic texture that overprints a clearly perceptible gneissic foliation defined by drawn-out K-feldspar and quartz grains, acicular sillimanite and flattened garnet porphyroblasts (Alₙ₉₋₇₁Prₚ₂₁₋₂₅Grₛ₂). Notable reaction textures in the rocks relate to
Figure 2. Variations in chemical parameters of minerals within anorthosite-leuconorite massif, ferrodiorite at massif margins and the bordering felsic rocks. Core and rim compositions of primary minerals shown by filled and open circles respectively; open triangle refers to core compositions of symplectitic and/or coronal minerals.

the stabilization of cordierite ($X_{Mg} \approx 0.62–0.69$) at the expense of garnet and sillimanite ($2Alm + 4Sil + 5Qtz = 3Fe-Crd$), and the replacement of garnet by intergrowths of cordierite + biotite + quartz ($6Alm + 4Kfs + 4H_2O = 3Ann + 3Fe-Crd + 3Qtz$ and $9Alm + 3Sil + 5Sa + 5H_2O = 6Crd + 5Ann$) and biotite + sillimanite ($2Alm + 2Kfs + 2H_2O = 2Ann + 2Sil + 4Qtz$).

Mafic granulites display granoblastic textures that often obliterate the crudely perceptible gneissic foliation in the rocks. The rocks contain orthopyroxene ($X_{Mg} \approx 0.49$), clinopyroxene ($X_{Mg} \approx 0.67$), plagioclase ($An_{48}$), hornblende, ilmenite and rare apatite, quartz and K-feldspar. Garnets ($Alm_{58}Prp_{15}Grs_{21}Sp_{6}$) occur strictly as rare coronas at the interface between pyroxene and plagioclase. By contrast, meta-gabbro/norites localized along the prominent shear zone in the southern part of the mapped area are massive and coarse-grained rocks. They display ophitic texture that is at places, obliterated by shear-induced protoclastic granulation.

Figure 2 depicts the mineral chemical variations in anorthosite-leuconite, ferrodiorite and quartz monzonite-granite gneisses. It is evident that the An content of plagioclase and $X_{Mg}$ of ferromagnesian phases decrease from anorthosite through leuconorite to ferrodiorite, and increase thereafter from quartz monzonite to granite gneisses. It is also important to note that the An content in plagioclase is similar in leuconorites and ferrodiorites.

4. Geochemistry

Forty samples comprising 9 anorthosite, 6 leuconorite, 2 ferrodiorite, 11 quartz monzonite gneiss, 6 granite gneiss, 2 megacrystic K-feldspar-bearing granite and 4 mafic granulite were analysed for major and selected trace elements by X-ray fluorescence, rare earth elements by neutron activation analysis and oxygen isotope ratios by mass spectrometry at the University of Bonn. Table 1 summarizes the mean and standard deviation of element abundances within each group. REEs are normalized to chondrite values after Taylor and McLennan (1985). Harker diagrams and element vs. element plots for the Turkel rocks are presented in figure 3. The samples from the Chilka Lake anorthosite complex (data from Sarkar et al 1981) are shown as fields in the figures for comparison.

Anorthosites and leuconorites show a calc-alkaline affinity, and are characterized by high Na$_2$O + K$_2$O (5.18 ± 0.43 wt.%), moderate to low mg$^+$ (16 to 42) and low abundances of plagioclase-incompatible elements. Sr (570 ± 79 ppm) and Ba (459 ± 79 ppm) contents are high. The REE patterns are typical for massif anorthosite with ($La/Lu$)$_{ch}$ between 7.75 and 75.50, and Eu* in the range 1.55 to 12.07. $\delta^{18}O$ values are clustered at 7.14 ± 0.18%. By comparison, ferrodiorites are poorer in Si, Al, Ca, Na and richer in Fe, Mn, Mg, Ti and P. The rocks are characterized by high Fe content (total Fe as Fe$_2$O$_3$ up to 32 wt.%), very low mg$^+$ (9 to 10) and are exceptionally rich in
Table 1. Mean compositions and standard deviations (in parenthesis) of major elements (in wt. %), selected trace and rare earth elements (in ppm) and $\delta^{18}$O values (in %) of igneous intrusives at Turkel.

<table>
<thead>
<tr>
<th>Rock Type</th>
<th>Rock</th>
<th>Anorthosite</th>
<th>Leuconorite</th>
<th>Ferrodiorite</th>
<th>Qtz-monzonite gneiss</th>
<th>Granite gneiss</th>
<th>Megacrystic granite</th>
<th>Mafic Granulite</th>
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<tr>
<td></td>
<td>(n)</td>
<td>(9)</td>
<td>(6)</td>
<td>(2)</td>
<td>(11)</td>
<td>(6)</td>
<td>(2)</td>
<td>(4)</td>
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<td>SiO$_2$</td>
<td>54.00 (0.69)</td>
<td>52.76 (1.17)</td>
<td>42.30 (4.12)</td>
<td>3.41 (2.79)</td>
<td>67.82 (3.05)</td>
<td>72.47 (4.41)</td>
<td>48.12 (1.13)</td>
<td></td>
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<tr>
<td>TiO$_2$</td>
<td>0.26 (0.15)</td>
<td>0.85 (0.40)</td>
<td>2.38 (0.15)</td>
<td>0.72 (0.18)</td>
<td>0.73 (0.29)</td>
<td>0.33 (0.18)</td>
<td>1.32 (0.78)</td>
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<tr>
<td>Al$_2$O$_3$</td>
<td>27.19 (0.78)</td>
<td>24.75 (2.64)</td>
<td>13.37 (2.75)</td>
<td>15.26 (0.82)</td>
<td>14.88 (0.67)</td>
<td>14.64 (1.42)</td>
<td>11.36 (2.62)</td>
<td></td>
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<tr>
<td>Fe$_2$O$_3$</td>
<td>1.59 (0.97)</td>
<td>5.37 (3.35)</td>
<td>27.61 (6.80)</td>
<td>7.70 (2.21)</td>
<td>5.35 (1.93)</td>
<td>2.32 (1.14)</td>
<td>15.28 (2.05)</td>
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<td>MnO</td>
<td>0.02 (0.01)</td>
<td>0.06 (0.04)</td>
<td>0.28 (0.03)</td>
<td>0.10 (0.04)</td>
<td>0.07 (0.03)</td>
<td>0.04 (0.00)</td>
<td>0.19 (0.03)</td>
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<tr>
<td>MgO</td>
<td>0.39 (0.35)</td>
<td>0.92 (0.34)</td>
<td>1.41 (0.21)</td>
<td>0.30 (0.09)</td>
<td>0.73 (0.32)</td>
<td>0.39 (0.10)</td>
<td>10.84 (3.64)</td>
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<tr>
<td>CaO</td>
<td>10.52 (0.35)</td>
<td>9.89 (1.37)</td>
<td>7.86 (0.73)</td>
<td>3.19 (0.59)</td>
<td>3.20 (0.85)</td>
<td>2.30 (0.52)</td>
<td>11.28 (1.91)</td>
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<tr>
<td>Nb</td>
<td>3.99 (0.26)</td>
<td>3.60 (0.31)</td>
<td>2.26 (0.36)</td>
<td>2.37 (0.09)</td>
<td>2.02 (0.12)</td>
<td>2.11 (0.36)</td>
<td>1.38 (0.75)</td>
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<tr>
<td>Yb</td>
<td>1.34 (0.13)</td>
<td>1.35 (0.29)</td>
<td>1.64 (1.03)</td>
<td>6.51 (0.63)</td>
<td>5.05 (0.81)</td>
<td>5.14 (1.25)</td>
<td>0.16 (0.10)</td>
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<tr>
<td>Total</td>
<td>99.46 (0.65)</td>
<td>99.82 (0.59)</td>
<td>100.30 (0.28)</td>
<td>99.79 (0.43)</td>
<td>100.08 (0.47)</td>
<td>100.06 (0.06)</td>
<td>100.03 (0.31)</td>
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$\delta^{18}$O values (in %) are 7.2 (0.1) for Anorthosite, 7.0 (0.3) for Leuconorite, 6.9 for Ferrodiorite, 8.6 (0.4) for Qtz-monzonite gneiss, 10.5 (0.2) for Granite gneiss, 18.0 (0.4) for Megacrystic granite, and 10.5 (0.2) for Mafic Granulite.

Mafic granulites are subalkalic tholeiites having low to intermediate Mg# (46–67) and mantle-like $\delta^{18}$O values (5.4 ± 1.3‰); REE patterns (20X chondrite) are flat with negligibly small Eu anomaly.

5. Petrogenetic considerations

A perusal of figures 3 and 4 show that the anorthosite, leuconorites and ferrodiorites describe fairly continuous curvilinear trends. But, the high-K felsic rocks lie off the trends. The quartz monzonite gneisses and the granitic rocks plot in distinct compositional fields that share only a minimum overlap. In the variation diagrams, the high-K felsic suite and ferrodiorites appear to plot along linear trends. But the two suites are separated by a large compositional gap and, therefore, these chemical lineages do not reflect magmatic differentiation trends. These observations point to the fact that the high-K felsic suite and the low-K anorthosite-leuconorite-ferrodiorite suite do not share a common parentage. The non-consanguinity of the suites is also reflected in the large differences in whole rock $\delta^{18}$O values from the high-K and low-K suites. Published $\delta^{18}$O values for melt-mineral pairs in basaltic systems are mildly negative...
 (>−0.4‰) or close to zero for plagioclase and mildly positive (1.0 to 0.3‰) for pyroxenes in the temperature range 1000–1200°C (Kyser et al 1982; Muehlenbachs and Kushiro 1974. Therefore, closed system equilibrium crystallization of plagioclase-dominated cumulates (plagioclase ≫ pyroxenes) would mildly enrich the residual magma in the heavier isotope, and even lead to a small decrease in δ¹⁸O values of the residual melts. Accordingly, a mantle-derived basic magma with δ¹⁸O ≈ 6‰ would yield a residual melt having similar or even lower δ¹⁸O values. Thus, the δ¹⁸O characteristics of felsic intrusives bordering the massif, if they were residual melts of anorthosite crystallization, cannot be explained by fractionation of plagioclase-dominated solids. Furthermore, the ubiquitous occurrence of garnet porphyroblasts in the quartz monzonite and granite gneisses and their absence in the anorthosite-leuconorite are difficult to explain if the rocks are deemed to be comagmatic. It is important to note that garnets in felsic rocks are, chemically and texturally informed to be of crustal origin and, therefore, are not liquidus phases.

The above arguments, when considered along with the mineral-chemical variations in the rocks shown in figure 2, prove that the anorthosite-leuconorite suite and the K-rich felsic suite are not genetically related. This appears to be true also for the Chilka Lake samples (cf. Sarkar et al 1981), although the element trends described by the Chilka Lake anorthosite-leuconorites are slightly shifted compared to their counterparts at Turkel. It stands to reason, therefore, that the comagmatic nature of the low-K anorthosite-leuconorite suite and the high-K quartz monzonite-granite suite as advocated for the Chilka Lake by Sarkar et al (1981) and tacitly assumed for the Eastern Ghats massif anorthosite complexes in general by Bose (1979) and Leelanandam and Narsimha Reddy (1988) is not tenable. It also follows that the validity of the isochron determined by Sarkar et al (1981) for the Chilka Lake complex is open to scrutiny.

The overall composition of the anorthosite-leuconorite-ferrodiorite suite may be calculated by multiplying the areal extent of the independent members
in the suite (anorthosite 62%, leuconorite 37%, ferrodiorite ~1% at the present level of erosion) with their respective bulk compositions (average values in table 1). Making some allowance for the presence of cumulus ferromagnesian phases (dominantly pyroxenes) that possibly reside at deeper levels (unexposed), the composition of the magma parental to the anorthosite suite is determined to be extremely alluminous, of intermediate mg# and with SiO₂ close to 50 wt.%. The abundance of large-ion lithophile elements (e.g., K and Ba) and transition elements (e.g., V, Cr and Ni) is very low. δ¹⁸O is ~7‰. REEs are low in abundance and characterized by LREE-enriched weakly fractionated patterns with positive Eu anomaly. These chemical features indicate that the magma parental to the anorthosite suite was akin to fractionated basic melts, presumably resembling high-Al gabbros as has been suggested by Mitchell et al (1996) for the Laramie anorthosite complex. However, field evidence in support of the conclusion is lacking.

By contrast, the felsic intrusives were presumably derived by incongruent melting of crustal rocks. The following chemical features of granite gneisses and megacrystic K-feldspar-bearing granites are suggestive of crustal origin, e.g., the peraluminous nature, high contents of SiO₂, K₂O and Ba, high K₂O/Na₂O and Rb/Sr ratios, the high δ¹⁸O values and high Fe#. The ensemble of minerals enclosed in restitic garnets in the rocks also support the above contention. Granitic rocks, texturally and chemically similar to those reported here, occur throughout the Eastern Ghats Belt as megaboudins and syn-tectonic massifs, exclusive of massif anorthosites. Mukhopadhyay (1995) has interpreted similar rocks in the Salur massif to have been derived by a high degree of incongruent melting of biotite-bearing metagreywacke gneisses followed by restite-melt unmixing. Isotope
Figure 4. Rare earth and trace element variations in anorthosite, leuconorite, ferrodiorite, quartz monzonite gneiss and mafic granulites at Turkel. Symbols as in figure 3.

Figure 5. Pseudoternary diagram after Roedder (1979) showing the fields 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). Note the compositional gap between the Fe-rich ferrodiorites (solid circles) and Si-rich quartz monzonite (triangles)-granitic rocks (inverted triangles) at Turkel. Field for high-Zr ferrodiorites and spatially associated felsic rocks from Laramie after Mitchell (1996) shown as open field.

studies on the rocks primarily from Salur yield whole rock $\varepsilon^{144aNd}(Nd) \approx -6$ to $-12$, $\varepsilon^{140Sr}(Sr) \approx 100$ to 1000, and for leached K-feldspars $^{206}Pb/^{204}Pb \approx 38$, $^{207}Pb/^{204}Pb \approx 15.9$ and $^{208}Pb/^{204}Pb \approx 18$ (cf. Krause et al. 1996). Clearly, the isotope data prove a crustal origin for the rocks.

The metaluminous quartz monzonite gneisses are higher in $K_2O$, $Ba$, $K_2O/Na_2O$ and $fe^\#$, but lower in $Rb/Sr$ compared to the granitic rocks. The Zr content in the rocks is far in excess compared to the Zr saturation levels experimentally determined in melts of comparable major element chemistry (Watson and Harrison 1983, 1984). It is obvious that the quartz monzonite melts inherited zircon either from extraneous crustal sources or from crustal precursors that produced the quartz monzonites on melting. The second possibility appears more realistic because zircons are found to be enclosed within restitic garnets. However, the metaluminous nature of the quartz monzonite argues against the rocks being S-type granitoids. This is also partly supported by the low $\delta^{18}O$ values in the rocks. It is possible that the quartz monzonites were derived by a high degree of incongruent melting of metamorphosed volcano-sedimentary piles or even tonalite gneisses.

5.1 The origin of ferrodiorites

The unusual chemistry of the Turkel ferrodiorites, e.g., the unusually high Fe content ($Fe_2O_3$ up to 32 wt.%), the exceptionally high $fe^\#$ ($\approx 91$) and the high abundance of high-field strength elements (Zr up to 4674 ppm and REE up to 150X chondrite), is typical for extremely fractionated melts. In literature, ferrodiorites are deemed to be either segregated Fe-rich melts formed due to unmixing of Fe-rich (ferrodiorite ?) and Si-rich (quartz monzonite ?) liquids or residual melts of anorthosite crystallization. It is, therefore, imperative to examine which of the two processes was responsible for the formation of ferrodiorites.

Experimental data on two-liquid partition coefficients after Watson (1976), Ryerson and Hess (1978) and Ellison and Hess (1991) in synthetic silicate melts of basic composition show that (a) $FeO$, $MnO$, $MgO$, $TiO_2$, $CaO$, $P_2O_5$ among major elements are partitioned into the Fe-rich liquid, whereas, $SiO_2$, $Al_2O_3$, $Na_2O$ and $K_2O$ are enriched in the Si-rich melts, (b) Mg/Fe ratios in the unmixed melts are similar, (c) Sr, Ba, Rb, Zr and REEs are strongly enriched in the Fe-rich liquids, and (d) REEs are preferentially partitioned into the Fe-rich liquids, but two-liquid partition coefficients are identical for all REEs. Figure 5 shows that the ferrodiorites and quartz monzonites plot in two distinct fields close to the 2-liquid solvus in quartz-fayalite-leucite pseudoternary. Also, the relative abundances of most trace and rare earth elements in ferrodiorites and quartz monzonites broadly conform with the experimental results. Thus, at a glance, it appears that the two rocks formed due to liquid unmixing. However, the following discrepancies are evident. First, contrary to the experimental results, the Ba content in ferrodiorites ($602 \pm 446$ ppm) is much lower compared to quartz monzonites ($1502 \pm 303$ ppm). Second, $D^{18}O$ (cf. Ellison and Hess
Table 2. Mineral-melt (basic) $K_D$ values, modal mineralogy of assumed anorthosite-leuconorite solids, and initial compositions of anorthosite parental magma. See text for discussion.

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<th>$K_D$ values</th>
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<tr>
<td>La</td>
<td>0.180</td>
<td>0.026</td>
<td>0.288</td>
<td>0.000</td>
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<tr>
<td>Eu</td>
<td>0.340</td>
<td>0.099</td>
<td>0.354</td>
<td>0.000</td>
</tr>
<tr>
<td>Sr</td>
<td>3.600</td>
<td>0.050</td>
<td>0.300</td>
<td>0.000</td>
</tr>
<tr>
<td>Zr</td>
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<td>0.020</td>
<td>0.420</td>
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Modal mineral content of the assumed solids

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<th>B</th>
<th>C</th>
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<td>P1 Opx</td>
<td>0.95</td>
<td>0.91</td>
<td>0.85</td>
<td>0.81</td>
<td>0.70</td>
<td>0.65</td>
</tr>
<tr>
<td>Cpx</td>
<td>0.03</td>
<td>0.05</td>
<td>0.10</td>
<td>0.12</td>
<td>0.20</td>
<td>0.22</td>
</tr>
<tr>
<td>Ilm</td>
<td>0.01</td>
<td>0.02</td>
<td>0.03</td>
<td>0.04</td>
<td>0.06</td>
<td>0.08</td>
</tr>
</tbody>
</table>

Composition of anorthosite parent magma

<table>
<thead>
<tr>
<th></th>
<th>Lacn</th>
<th>Eucn</th>
<th>Sr</th>
<th>Zr</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>18.77 (10.19)</td>
<td>13.90 (4.02)</td>
<td>463 (35)</td>
<td>37 (22)</td>
</tr>
</tbody>
</table>

Values within parenthesis are standard deviations.

Figure 6. Calculated Zr vs. Sr of coexisting solids (broken lines) and liquids (solid lines) at different F values (in numbers). See text for discussion and table 2 for mineral-melt $K_D$ values, initial melt composition and modal mineralogy of assumed solids. Measured element abundances in anorthosite, leuconorite, ferrodiorite and mafic granulite shown for comparison (symbols as in figure 3). Subscripts s and l refer to solids and liquids respectively. The field for high-Al gabbros after Mitchell et al (1995) shown by dotted line.

1991) of REEs for ferrodiorites and quartz monzonites (assuming both to be ‘pure’ melts) increases systematically from 1.2–1.4 for La and Ce, ≈ 2 for Nd and Sm and 2.2–2.5 for Yb and Lu. Third, the small positive Eu anomaly in some of the quartz monzonites is not reflected in ferrodiorites indicating thereby that
either the partitioning of Eu is different from the trivalent REEs or else the two melts do not represent unmixed liquids in equilibrium. Furthermore, the absence of garnet porphyroblasts in ferrodiorites as opposed to their persistent presence in quartz monzonites is difficult to be explained if the Fe-rich ferrodiorite and the Si-rich quartz monzonite formed due to liquid unmixing. Thus, although appealing, the ferrodiorites are possibly late-stage melts that are residual to extensive anorthosite-leuconorite crystallization. This is supported by the progressive enrichment of plagioclase-incompatible elements from anorthosite through leuconorite to ferrodiorite and the element trends described by the suite shown in the variation diagrams.

In this context, two significant chemical aspects in the ferrodiorites need to be examined. These are (i) the absence of a negative Eu anomaly in ferrodiorites complementary to the pronounced positive Eu anomaly in anorthosites and leuconorites, and (ii) the unexpectedly high Sr content in ferrodiorite melts residual after extreme fractionation of plagioclase that are known to preferentially incorporate Sr.

An attempt has been made to explain these ambiguities by closed-system multistage crystal fractionation using the total equilibrium equation after Shaw (1970). Mineral-melt (basic) distribution coefficients of relevant minerals are adopted after the NEWPET software (table 2). The initial element concentrations of the anorthosite-leuconorite-ferrodiorite parental magma are obtained by averaging the high-Al gabbros considered to be parental to the massif anorthosites in the Grenville Province after Mitchell et al (1995), e.g., Sr=463±35 ppm, Zr=37±22 ppm, La<sub>cn</sub>=18.77±10.19 and Eu<sub>cn</sub>=13.90±4.02 (table 2). Using these values as the initial melt composition, element concentrations in melts and coexisting anorthosite solid (composition A in table 2) are calculated for equilibrium crystallization in the range 5 to 95% i.e. F (fraction of liquid remaining) between 0.95 and 0.05 respectively. The exercise is repeated several times taking the residual melt at 30% crystallization as the initial melt composition for the next set of calculations using anorthosite-leuconorite.

Figure 7. Calculated Eu<sub>cn</sub> and La<sub>cn</sub> of coexisting solids (broken lines) and liquids (solid lines) at different F values (in numbers). See text for discussion and table 2 for mineral-melt K<sub>D</sub> values, initial melt composition and modal mineralogy of assumed solids. Measured element abundances in anorthosite, leuconorite, ferrodiorite and mafic granulite shown for comparison (symbols as in figure 3). Subscripts s and l refer to solids and liquids respectively. The field for high-Al gabbros after Mitchell et al (1995) shown by dotted line.
solid assemblages modified to more basic compositions (B, C, D, E and F in table 2). In figures 6 and 7, the computed variations of Sr (compatible) vs. Zr (incompatible) and La_{an} vs. Eu_{an} (both incompatible, but mineral-melt $K_{D, Eu} > K_{D, La}$) in successive anorthosite-leuconorite solids and coexisting melts are compared with the measured abundances in the rocks.

It is evident from figure 6 that the measured abundances of Zr in ferrodiorite are approached only at unrealistically high degrees of crystallization ($F < 0.01$), but the computed Sr contents in the assumed solids and the coexisting melts at the high degrees of crystallization are much lower compared to the measured abundances in the anorthosite-leuconorite solids and ferrodiorite melts. The computations were repeated for several sets of Zr and Sr values in the initial melt. It was observed that the measured Sr content in anorthosites and leuconorites agrees favourably with the computed Sr abundance in the assumed solid for Sr = 600 ppm in the initial melt. But the Sr abundance in the melts in equilibrium with the leuconorite assemblages E to F at extremely high degree of crystallization is < 50 ppm, and much lower than the measured abundance in ferrodiorites, e.g., ≈ 250 ppm. It is obvious, therefore, that the ferrodiorites are not 'pure' residual melts, and instead are mixtures of residual melts and plagioclase crystals inherited from the host solids in equilibrium with the melt. This is supported by the overlap of the An content in plagioclases in ferrodiorites and leuconorites. If the plagioclase crystals in ferrodiorites were precipitated from melts that are residual after leuconorite crystallization, then the An content in the ferrodiorite plagioclases should be lower than in the leuconorites.

There are two ways in which the computed Zr content in ferrodiorite may be increased to the desired level, i.e., either by revising the assumed Zr content in the initial melt to higher values or by selective dissolution of Zr in the residual melt. Computations using Zr = 150 ppm in the initial melt show that, at high degree of crystallization ($F = 0.05$), the Sr content in the assumed anorthosite solids (assemblages A and B) and leuconorite solids (assemblages C to F) is in the range 34 to 58 ppm and 96 to 407 ppm respectively. These values, albeit on the higher side, are comparable to the measured Sr abundance in anorthosites and leuconorites. But the computed Zr content in the corresponding residual melts in equilibrium with the assumed anorthosite-leuconorite solids are in the range 2360–3137 ppm and 4170–9136 ppm respectively. These values are too high compared to the measured Zr content in the ferrodiorite. Clearly, Zr value (90 to 120 ppm) in the initial melt will suffice.

In the alternate scenario, the residual Fe-rich melt following extraction from the host anorthosite-leuconorite solids is likely to come in contact with the zircon-bearing quartz monzonite melts bordering the massif. Since zircon is heavier than the quartz monzonite melt, some zircon grains are expected to settle at the interface between the siliceous melt and the underlying Fe-rich residual melt. These zircon grains are likely to be dissolved in the Fe-rich residual melts because such melts are known to be grossly undersaturated with respect to elemental Zr (Watson and Harrison 1983). This would result in the redistribution of elemental Zr and zircon-hosted elements like Y, Nb and HREE over LREE in the Fe-rich melt. It is understandable that only a limited amount of zircon assimilation is necessary to cause a large spike of Zr in the small amount of residual melts. Someapatite may also be similarly assimilated. However, it is important to note that the strong gradients in composition across the interface of the two contrasting melts are likely to be preserved because element exchange across the 2-liquid interface shall presumably be in accordance with similar laws that govern element partitioning in immiscible melts.

In figure 7, La_{an} and Eu_{an} values for the computed solids, at high degree of crystallization, match the measured element concentrations in the anorthosites and leuconorites. But the computed La_{an} and Eu_{an} values in the coexisting melts are much higher than the measured abundances in ferrodiorites. It is evident, however, that the ferrodiorites plot on a mixing line joining the computed element concentrations in the coexisting solids and liquids at a high degree of crystallization. This implies that a suitable mixture of computed anorthosite-leuconorite solids and late-stage melts would adequately explain the REE abundances in the ferrodiorites. Or, in other words, the ferrodiorites may be considered to be residual melts that contain in suspension plagioclase crystals inherited from the solids in equilibrium. This will cause, depending on the amount of uptake of plagioclase crystals by the melt, a marked decrease in the La content and a less marked decrease in the Eu content in the late-stage melts. The net effect would be to suppress any negative Eu anomaly that may be present in the residual melts.

The above considerations, taken together, prove that the trace and rare earth element chemistry of ferrodiorites corresponds to late-stage melts that are residual to extensive anorthosite-leuconorite crystallization from high-Al gabbroic melts, but the residual melts were not entirely segregated from the host anorthosite-leuconorite. It is, however, important to note that in the adjacent anorthosite complex at Bolangir (Raith et al. 1997, this volume), the ferrodiorite residual melts were completely segregated from the anorthosite-leuconorite solids. This is reflected in the lower An content in plagioclase phenocrysts in the Bolangir ferrodiorites compared to the coexisting solids.
Continued crystallization of plagioclase + pyroxene along the boundary curve caused olivine to reappear at the olivine + pyroxenes + plagioclase coticetic. At the end stage of this process, the residual melt became highly enriched in Fe and plagioclase-incompatible elements. Petrographic observations indicate thatapatite and zircon saturation limits in the melts were eventually reached. The dense ferrodiorite melts supported by the densely-packed plagioclase crystal much were subsequently filter-pressed into 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 the filter-pressed Fe-rich melts to inherit plagioclase crystals from the coexisting leucorinite members.

The high-K felsic intrusives bordering the massif are restitic garnet-bearing melts generated by incongruent melting 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 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-leucorinite diapir rising in their wake en route to the final emplacement of the igneous intrusives.

Acknowledgements

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6. The magmatic evolution of the Turkel anorthosite complex

Extensive fractionation of early crystallizing ferromagnesian phases (olivine and low-Ca pyroxenes) caused mantle-derived basic magmas to become enriched in Al over Si, Fe over Mg and in the plagioclase component. The ascent of these fractionated basic magma bodies and progressive Fe-enrichment due to continued fractionation of ferromagnesian silicates in the ascending magmas caused the melt composition to move towards the expanding plagioclase liquidus surface (Longhi and Pan 1988; Fram and Longhi 1992; figure 8). This in turn caused the melt composition (high-Al gabbro) to become stranded within the plagioclase + melt field. This led to voluminous precipitation of plagioclase in the magma pools ponded below a continental lid. The buoyant plagioclase-rich crystal mush during ascent trapped pools of resident magmas. Decompression in the trapped magma pools led to more plagioclase crystallization, forcing the residual melt composition to move away from the plagioclase apex towards the pyroxene + plagioclase boundary curve (figure 8).
Turkel anorthosite complex


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