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THE TORA TIGGA COMPLEX, SOUTHERN DIR, NW PAKISTAN; AN EXAMPLE OF MAFIC-ULTRAMAFIC ROCKS IN THE BOTTOM OF AN ISLAND ARC

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ABSTRACT

The Tora Tigga complex comprises dunite, peridotites and pyroxenites (± hornblende), hornblendites, metagabbros and plagiogranite covering over 6 km^2 area. It is located in the north of the Indus-Zangbo suture, in an extensive terrain of amphibolites that form the southern part of the Cretaceous Kohistan island arc. The ultramafic rocks are closer to those found in stratiform complexes than to oceanic peridotites and the metagabbros are calc-alkaline island arc-type. Chemical variation suggests that the rocks are magmatically related. They probably represent amphibolite facies metamorphosed cumulates in a magma chamber of the Kohistan arc. The hornblendites are medium-grained to pegmatitic, and range from small dykes, patches and pools to large bodies over 2.5 km in length. There are gradational lithologies to pyroxenites and peridotites in which hornblende is replacive in texture; however, chemical data do not support a metasomatic origin for such rocks and the hornblendites.

INTRODUCTION

Northern Pakistan is characterized by the presence of two suture zones, the Main Karakoram Thrust (MKT) in the north and the Main Mantle Thrust (MMT) in the south. These are the westerly bifurcation of the Indus Zangbo suture and enclose ~ 36,000 km^2 area of the Cretaceous Kohistan island arc. North of MKT lies the Karakoram plate and south of MMT the Indo-Pakistan plate. the Kohistan arc is tilted so that a complete cross section is exposed. Along a N to S traverse across the arc, the following major lithologies, each stretching for several hundred kilometers, are observed: Middle Cretaceous Yasin sediments; Late Jurassic-Cretaceous Chalt volcanics; Creto-Tertiary calc-alkaline granitic belt; Early to Middle Cretaceous Chilas complex; and southern amphibolite belt (for further detailes, see Jan, 1977; Tahirkheli & Jan, 1979; Tahirkheli et al., 1979; Butt et al., 1980; Andrews-Speed & Brookfield, 1982; Coward et al., 1982, 1986; Bard, 1983; Petterson & Windley, 1986).

The Chilas complex comprises norites, with subordinate anorthosites, quartz-hypersthene diorites, gabbros, troctolites, pyroxenites, peridotites and dunites. These rocks may partly represent cumulates in the magma chamber of the arc and have re-equilibrated under pyroxene granulite facies conditions, followed by amphibolite facies retrogression during ascent (Bard, 1983; Jan et al., 1984). The southern amphibolite belt lies just north of the MMT and contains homogeneous and banded amphibolites derived from island arc volcanic and plutonic rocks that may be the metamorphic equivalent of the Dras volcanics found to the east of Nanga Parbat in Ladakh (Honegger et al., 1982). The belt also contains a variety of other lithologies such as metamorphosed gabbros/norites, troctolites, ultramafites (some of which resemble those of the Chilas complex), granitic rocks, and minor calcareous and siliceous metasediments.

The Tora Tigga complex (34° 49' N, 71° 44' E) is an example of the plutonic masses found in the amphibolite belt. It is located immediately north of the MMT and is characterized by an abundance of hornblendites. Based on a study of 170 thin sections and 23 analyses (major and trace elements by atomic absorption and XRF), a summarised petrographic and geochemical account of the complex is presented in this paper. Additional details can be found in Banaras & Ghani (1982), Jan et al. (1983) and Tahirkheli (1983).

PETROGRAPHY

The Tora Tigga complex consists of olivine- and pyroxene-rich rocks (henceforth referred to as ultramafic rocks), hornblendites, metamorphosed gabbros/norites, and plagiogranite, emplaced in a thick sequence of amphibolites. The ultramafic rocks and hornblendites occur in two exposures, 5.5 km^2 and 0.25 km^2 in area, which are separated by the host amphibolites. The southern limit of the complex is marked by the Jandul stream.

Amphibolites and metagabbros

The host amphibolites in the immediate vicinity of the complex have metaplutonic aspects. They contain enclaves of fine-grained amphibolite, and are composed of plagioclase (An_{45}), hornblende, garnet (up to 3 cm porphyroblasts) and quartz, with minor amounts of epidote, opaque oxide, apatite, sphene, rutile, biotite and chlorite. On the south-east these garnet amphibolites are separated from the metagabbro by a 100 m thick shear zone of banded amphibolites. These are devoid of garnet but contain more biotite; it is not clear whether they have developed from the garnet amphibolites or metagabbros. Plagioclase in these rocks is a little less calcic (An_{38}).

The metamorphosed gabbros/norites consist of plagioclase (An_{40-43}), orthopyroxene (up to 16%) and clinopyroxene (up to 5%) relics enclosed in hornblende, with quartz, opaque oxide and minor epidote. These rocks follow the pattern of Karroo dolerites on Niggli diagrams (cf. Leake, 1964). Lack of sedimentary features or associated sediments and the presence of medium-grained pyroxene relics, minor anorthositic bands and thin doleritic dykes suggest that they are metamorphosed gabbros/norites. The occurrence of anorthosite and mafic dykes, and the presence of strongly pleochroic orthopyroxene and greenish clinopyroxene are typically reminiscent of the norites of the Chilas complex (Jan & Howie, 1980; Jan et al., 1984). Because noritic rocks of the Chilas complex occur to the N (Kakar et al., 1971) and E (Chaudhry et al., 1974), it is possible that the Tora Tigga metagabbroic rocks are an extension of the Chilas complex (Jan et al., 1983).

Ultramafic rocks

These include dunite, peridotites and pyroxenites which contain variable amounts of a hornblendic amphibole. Some of them show deformational features such as granulation and kink bands, and all are variably altered along at least three sets of fractures. Serpentine, magnetite and, in some rocks, talc are the alteration products, developed more readily after olivine than pyroxene.

The dunite contains over 90% olivine, the remainder being orthopyroxene (locally poikilitic) and chromite. the periodtites contain one or two pyroxene, minor opaque oxide and over 70% olivine. In the amphibole peridotites, olivine and serpentine make 45 to 60%, ortho and clinopyroxene 1 to 15% each, and hornblende 7 to 40%; opaque oxide and, rarely, secondary carbonate are subordinate. These rocks, lumped together as olivine ultramafites in Fig.1, form outcrops generally 300 m in length.

The pyroxene-rich ultramafites vary in texture and mineralogy, and can be classified into (1) hornblende websterite, the most common, passing into (2) hornblende orthopyroxenite, (3) hornblende-olivine websterite, (4) olivine clinopyroxenite, (5) clinopyroxenite, the least common and passing into (6) hornblende clinopyroxenite. The rocks contain opaque oxide but are devoid of plagioclase. Secondary serpentine, magnetite, talc and chlorite replacement took place after the development of hornblende. The six types could not be separately mapped. The ultramafite bodies extend E-W with intervening metagabbros and hornblendites; the largest body is up to 400 m broad and 2 km long. The rocks contain many patches and veins of hornblende and their contacts with hornblendites are gradational.

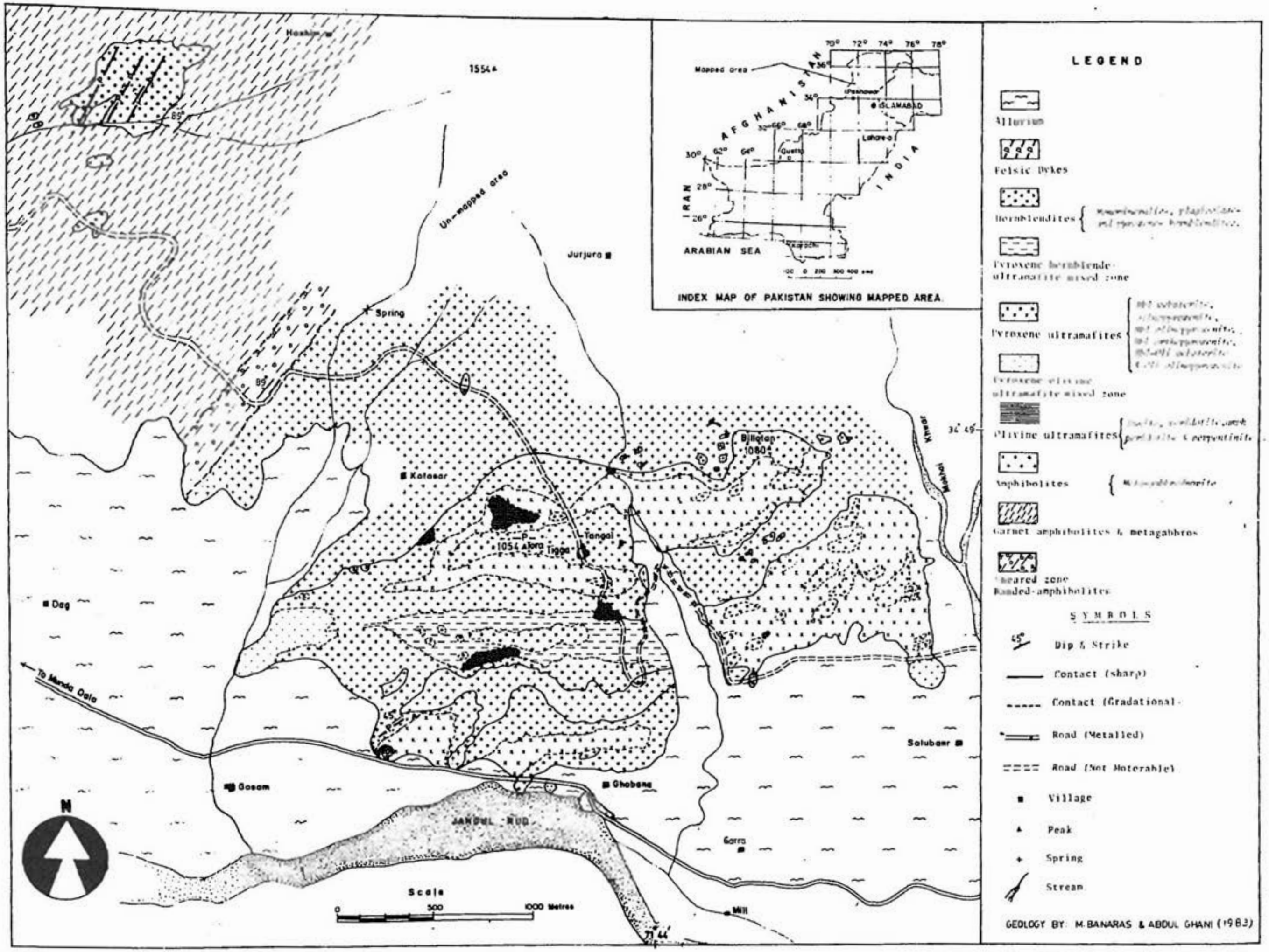


Fig. 1. Geological map of the Tora Tigga complex, Munda area, Dir.

The hornblende is colourless, green, or brownish green. It is post-cumulus and occurs interstitially and as "poikiloblasts" containing pyroxene and opaque oxide. The enclosed pyroxene grains mostly lack crystal boundaries and are finer grained than the independent pyroxene grains. Some orthopyroxene grains are surrounded by a girdle of magnetite granules probably representing excess iron over that accommodated in the growing hornblende poikiloblast. In rare cases, an orthopyroxene inclusion may be shared by more than one hornblende poikiloblasts. The hornblende has also grown along cleavages and fractures in the clinopyroxene, and in some orthopyroxene grains it occurs as blebs that may have grown at the expense of exsolved clinopyroxene. There are clinopyroxene grains with a spongy growth of hornblende that, in a few cases, is in optical continuity with neighbouring hornblende poikiloblasts. The poikiloblasts may rarely contain hornblende inclusions or they may be surrounded by smaller hornblende grains. These features are atypical of igneous rocks and, coupled with field data, may suggest a complex interplay of replacement and deformation (Jan et al., 1983).

Hornblendites

Forming the principal lithology of the complex, the hornblendite bodies reach up to 250 m in width and up to 3 km in length. There also are numerous veins, dykes, pools and patches, less than a meter to tens of meters, in the metagabbros and pyroxenites, and rare enclaves of the later two in the hornblendites. Some contacts are gradational but some, especially those of the Hashim body, are sharp or sheared, which can be explained by invoking subsequent deformation (Banaras & Ghani, 1982).

The rocks can arbitrarily be divided into monomineralic-, plagioclase- and pyroxene hornblendites. Those containing plagioclase (mostly in the Hashim body) are generally devoid of pyroxene and vice versa. The plagioclase (An_{45-50}) is generally 10% but reaches up to 25%. Clinopyroxene makes up to 35% and orthopyroxene, found only in some rocks, reaches up to 10%. The rocks contain minor amounts of magnetite, apatite, rutile, sphene, quartz, epidote, talc, chlorite and, rarely, carbonate and prehnite. The last-named five minerals and much of quartz are secondary in origin.

The hornblende is generally fresh or only slightly altered to epidote and chlorite, but the plagioclase is commonly saussuritized; both may contain magnetite and rutile inclusions. The hornblende is green or brownish green, with bluish green margins in some cases. Twinning is common and mostly multiple. It may be continuous, partial or edge type, but a single grain normally contains only one type. The orthopyroxene is changed over to talc, magnetite $\pm$ carbonate which are, in rare cases, separated from hornblende by a thin envelope of actinolitic amphibole. Textural relations between pyroxene and hornblende in pyroxene hornblendites are similar to those described for the hornblende pyroxenites with which they are closely associated and have gradational contacts.

The hornblendites, especially the monomineralic ones, are heterogeneous in grain-size and range from medium-grained to pegmatitic (up to 25 cm long crystals). Most, however, are coarse-grained with up to 3 cm long grains. The pegmatitic variety is monomineralic and in several places occurs as dykes and veins in finer-grained hornblendites. Deformational features are not obvious megascopically, but in thin section the hornblende may show granulation, kinking and bending. That the hornblendites must have passed through episode(s) of deformation is documented in the plagiogranite dykes that are folded, foliated and strongly cataclasized.

The boundaries of the hornblende grains are straight or slightly curved and they tend to form triple point equilibrium angles of 120°. Such angles are attributed to result from static balance of three equal interfacial tensions, annealing and recrystallization during metamorphism (Jackson, 1961; Vernon, 1968; Spry, 1969; Moore, 1973). Deviation to lobate, dentate, embayed and irregular boundaries are also present, the latter being more common in granulated rocks. In a number of cases, one hornblende grain may penetrate another, in some all the way through. There also are rare trails of medium-grained e hedral to subhedral hornblende grains passing through large grains. These may represent complex deformational and metamorphic features or hornblende of two generations--magmatic and metamorphic/metasomatic. More likely, the hornblende grains may have enlarged themselves by using most units of adjacent hornblende by replacement (cf. Spry, 1969), possibly after the earlier grains were fractured.

No hornblende was chemically analysed but five composition were calculated from monomineralic hornblendites after making suitable corrections for minor impurities of epidote, magnetite, sphene and rutile. According to the IMA nomenclature (Leake, 1978), one classifies as pargrastic hornblende, another as tschermakitic hornblende, and three as magnesiohastingsitic hornblende. Jamieson (1981) found that in the St. Anthony complex, metamorphic hornblendes have distinctly lower Na+K and Al^{iv} than igneous hornblendes. All the five calculated compositions from Tora Tigga fall in the field occupied by the metamorphic hornblendes of the St. Anthony complex.

Minor dykes and veins

Veins and patches containing one or more of the phases amphibole, epidote, plagioclase, chlorite, carbonate and quartz, probably developed during uplift, are common in the complex. Like elsewhere in the southern part of the Kohistan arc, there are rare plagioclase hornblende pegmatites of presumed metasomatic origin (Jan, 1977). The metagabbros, like those of the Chilas complex (cf. Jan et al., 1984) are also intruded sparsely by veins and dykelets of amphibolitic composition.

There are many leucroatic veins and up to 2 m thick dykes of plagiogranite. These consist essentially of plagioclase and quartz, with some amphibole and minor amounts of magnetite, apatite, chlorite and epidote. These are folded and deformed, with gneissose to mylonitic fabric. Some mylonites consist of over 70% groundmass, the remainder being porphyroclasts of both hornblende and plagioclase. The Hashim body is cut through by three parallel dykes each of which is about 1.5 m thick. In Tora Tigga, the plagiogranite dykes make parallel swarms trending E-W and dipping north.

GEOCHEMISTRY

Eight ultramafic rocks, nine hornblendites, five metagabbros and one plagiographite were analysed by XRF/atomic absorption. Twelve representative analyses are listed in Table 1. Details of analytical techniques are given in Tahirkheli (1983).

The ultramafic rocks contain low quantities of TiO_2, Al_2O_3, MnO, Na_2O, K_2O, and are akin to plagioclase-free cumulates in the stratiform complexes and harzburgite sub-type ophiolites. Lherzolite sub-type peridotites contain higher Al_2O_3 and Na_2O (Jackson & Thayer, 1972; Green, 1964). There is a considerable variation in the amounts of H_2O+ and Fe_2O_3 (both 1 to 7%), the two showing a rough positive correlation suggestive of the degree of serpentinization. Therefore, if the value of Fe_2O_3 is converted to FeO, then the mole ratio MgO/(MgO+FeO+MnO) can be more meaningfully compared for the various rocks. This ratio for residual ultramafic rocks in ophiolite complexes is about 0.91 (Coleman & Keith, 1971; Loney et al., 1971; Himmelberg & Loney, 1973; Moore, 1973; Miyashiro, 1975; Jan & Howie, 1981). Stratiform ultramafic rocks generally have lower values for this ratio (Jackson & Thayer, 1972; Skinner et al., 1978). Values for this ratio in Tora Tigga ultramafic rocks are also lower than those of ophiolitic ultramafics and range from 0.78 to 0.83 (with one value of 0.87).

The hornblendite analyses contain 45% SiO_2 except in two pyroxene-bearing samples. All are olivine-normative and six contain nepheline in norms. The analysed pyroxene hornblendites, and a pure hornblendite with lower CaO than the rest, are hypersthene-normative. Compared to the ultramafic rocks the hornblendites have lower MgO and Niggli mg, and higher Al_2O_3, TiO_2, and alkalis. The monomineralic and plagioclase hornblendites are similar in chemistry, but the latter contain a little higher alkalis. The two pryroxene hornblendites are lower in Al_2O_3, TiO_2 and alkalis than the rest; the orthopyroxene-bearing rock contains more MgO, FeO*, and lower CaO, and the clinopyroxene-bearing one is higher in CaO and lower in FeO, as expected from their mineral composition. Of the five metagabbros, three are quartz-normative and two are olivine-normative. The five rocks show considerable variations in the amount of Sr, Cr, Zr and Ni.

TABLE 1. REPRESENTATIVE ANALYSES OF ROCKS FROM TORA TIGGA

	TT7	TT69	TT123	TT84	TT9	TT17	TT96	TT134	TT71	TT23	TT78	TT14	TT42	TT161
SiO_2	40.30	40.70	48.22	48.49	54.47	41.98	44.68	43.95	45.59	49.50	48.57	49.93	56.22	68.76
TiO_2	0.00	0.00	1.02	0.61	0.00	2.04	0.87	1.50	1.15	0.57	0.73	1.19	1.21	0.20
Al_2O_3	0.76	0.64	4.66	1.42	1.40	15.73	10.21	13.51	8.23	7.56	17.98	15.83	15.68	17.62
Fe_2O_3	1.63	1.55	2.26	2.06	1.01	6.48	4.13	3.69	5.19	4.91	3.41	2.54	2.59	0.20
FeO	13.46	12.71	8.30	11.49	4.04	5.86	6.13	8.07	8.07	4.26	5.48	4.07	4.15	0.30
MnO	0.20	0.25	0.14	0.22	0.14	0.16	0.12	0.17	0.19	0.17	0.11	0.14	0.14	0.02
MgO	40.12	38.30	21.74	32.14	18.68	13.54	16.35	13.79	19.26	17.60	7.82	9.50	5.08	0.42
CaO	2.55	5.32	12.08	2.52	19.56	9.63	13.27	10.80	7.65	13.25	10.17	10.32	8.82	3.81
Na_2O	0.04	0.19	1.35	0.63	0.33	2.04	1.83	2.44	1.71	0.90	3.55	3.00	3.31	6.04
K_2O	0.01	0.00	0.08	0.03	0.01	0.20	0.21	0.31	0.14	0.02	0.20	0.08	0.51	0.25
P_2O_5	0.00	0.22	0.00	0.03	0.12	0.25	0.10	0.20	0.17	0.12	0.29	0.18	0.22	0.01
$H_2O^{\pm}$	—	—	—	—	—	2.10	2.12	1.57	2.63	1.06	1.69	3.21	2.06	2.37
Rb	5	—	3	—	3	3	5	1	3	—	—	1	—	23
Sr	16	—	39	—	38	86	129	116	64	44	551	382	490	730
Ba	60	—	22	—	—	93	71	—	—	—	53	59	60	24
Y	2	—	6	—	7	12	10	12	8	6	10	13	11	4
Ce	33	—	59	—	17	22	30	—	—	—	—	17	—	—
Ni	1174	714	340	496	210	100	194	62	436	198	16	82	9	—
Cr	5586	716	1035	2447	1571	38	924	153	938	814	118	208	75	30
Co	182	—	59	—	36	77	39	93	99	65	34	35	37	100
V	159	—	384	—	159	532	344	468	305	316	229	226	198	—
Zr	9	—	13	—	13	22	24	31	15	8	120	76	163	340
						C.I.P.W. NORMS								
Ne	0.00	0.42	0.00	0.00	0.00	0.00	4.66	4.24	0.00	0.00	0.00	0.00	0.00	0.00
Q	0.00	0.00	0.00	0.00	0.25	0.00	0.00	0.00	0.00	0.00	0.00	0.00	11.10	24.49
Or	0.06	0.00	0.47	0.17	0.06	1.20	1.27	1.86	0.86	0.12	1.19	0.50	3.06	1.50
Ab	0.37	0.80	11.41	5.34	2.84	17.67	7.25	13.15	14.86	8.22	30.52	26.49	28.66	52.32
An	1.87	0.91	6.47	0.95	2.30	33.86	19.43	25.40	14.76	16.44	33.11	28.12	26.96	19.29
Cpx	8.78	18.87	42.57	9.05	74.35	10.39	37.75	22.15	18.29	38.69	13.06	19.02	12.95	0.00
Hy	2.64	0.00	3.65	43.50	18.45	0.63	0.00	0.00	18.60	24.75	2.48	18.52	10.57	1.16
Ol	83.61	71.47	30.23	36.52	0.00	22.08	21.72	24.42	22.25	3.31	12.52	0.71	0.00	0.00
Mt	2.17	6.98	3.27	3.06	1.47	9.60	6.12	5.44	7.73	7.19	5.03	3.84	3.83	0.30
Il	0.00	0.00	1.94	1.16	0.00	3.97	1.69	2.89	2.24	1.10	1.42	2.37	2.34	0.39
Ap	0.67	0.52	0.00	0.07	0.27	0.56	0.22	0.44	0.39	0.27	0.64	0.41	0.50	0.02

Analysts: A. Z. Tahirkheli & M. Q. Jan

All the analyses were recalculated to 100%; excluding H_2O and including Cr_2O_3 in the ultramafic ones. In the more serpentinized rocks (TT7, 69, 84) the amount of Fe_2O_3 was reduced to TiO_2 + 1.50 % according to Irvine and Baragar (1971).

TT7: dunite; TT69: peridotite; TT123: hornblende websterite; TT84: hornblende orthopyroxenite; TT9: hornblende clinopyroxenite; TT17 and 96: "pure" hornblendites (Nb=0.5) ; TT134: plagioclase hornblendite; TT71: orthopyroxene hornblendite; TT23: clinopyroxene hornblendite; TT78 and 14: metagabbros (Nb=1); TT42: meta-quartz diorite; TT161: plagiogranite.

On MFA diagram and according to the scheme of Wright (1969), the plagiogranite classifies as calc-alkaline. The Rb vs SiO_2, Nb vs Y, Ta vs Yb, Rb vs Y + Mn, and Rb vs Yb+Ta plots of this rock and those of the quartz diorite (TT42) fall consistently in the field of volcanic arc granites (cf. Pearce et al., 1984). On the albite-orthoclase-quartz system at an assumed PH_2O of 2000 kg/cm^2, the granite plots near 760° isotherm (Tuttle & Bowen, 1958). Either the granite has been produced by partial melting during metamorphism or it represents residual liquid formed during fractional crystallization.

Major oxides variation

Oxides were plotted against MgO, solidification index of Kuno (1959), mafic index of Wager & Deer (1939), and differentiation index of Thornton & Tuttle (1960); Niggli values were also plotted against mg. In most cases the analyses display systematic variations and smooth trends; the gaps between metagabbros and ultramafic rocks are bridged by the hornblendite analyses.

The oxides and Niggli mg variations against differentiation index are shown in Fig.2. Al_2O_3, K_2O and Na_2O show a gradual increase with DI, whereas Fe_2O_3 + FeO, FeO, Niggli mg, and MgO decrease, the latter drastically. A high water content in the magma may have expanded the stability fields of pyroxene and hornblende, thus leading to enrichment of alumina in the residual melts. The CaO, MgO, Fe and SiO_2 plots of some ultramafic rocks show scatter due to variations in the amounts of orthopyroxene, clinopyroxene and olivine; however, for the rest of the rocks SiO_2 increases and CaO decreases with DI.

TiO_2 increases with DI in ultramafic rocks and honrblendites but decreases in metagabbros. This can be explained if it is assumed that the three groups of rocks are derived from the same magma. Early fractionation of Ti-poor phases such as olivine, chromite and two pyroxenes caused a steady enrichment of TiO_2 in the magma. During the following stages of fractionation, a lot of TiO_2 was withdrawn from the magma in hornblende, thus causing its decline in later-formed (meta) gabbroic rocks. Decrease in TiO_2 and iron oxide and increase of SiO_2 with differentiation are considered to be typical of calc-alkaline rocks (Miyashiro, 1975).

Trace element variation

Trace elements were plotted against mafic and solidification indices. In both cases, Cr, Ni, Sr, and Y show systematic variation. Ce, Zn and Ba plots are inconclusive, possibly because of limited data. Fig. 3 displays that Cr, Ni, V, and Co decrease and Sr and Zr increase systematically with differentiation index. Cr & Ni are withdrawn from

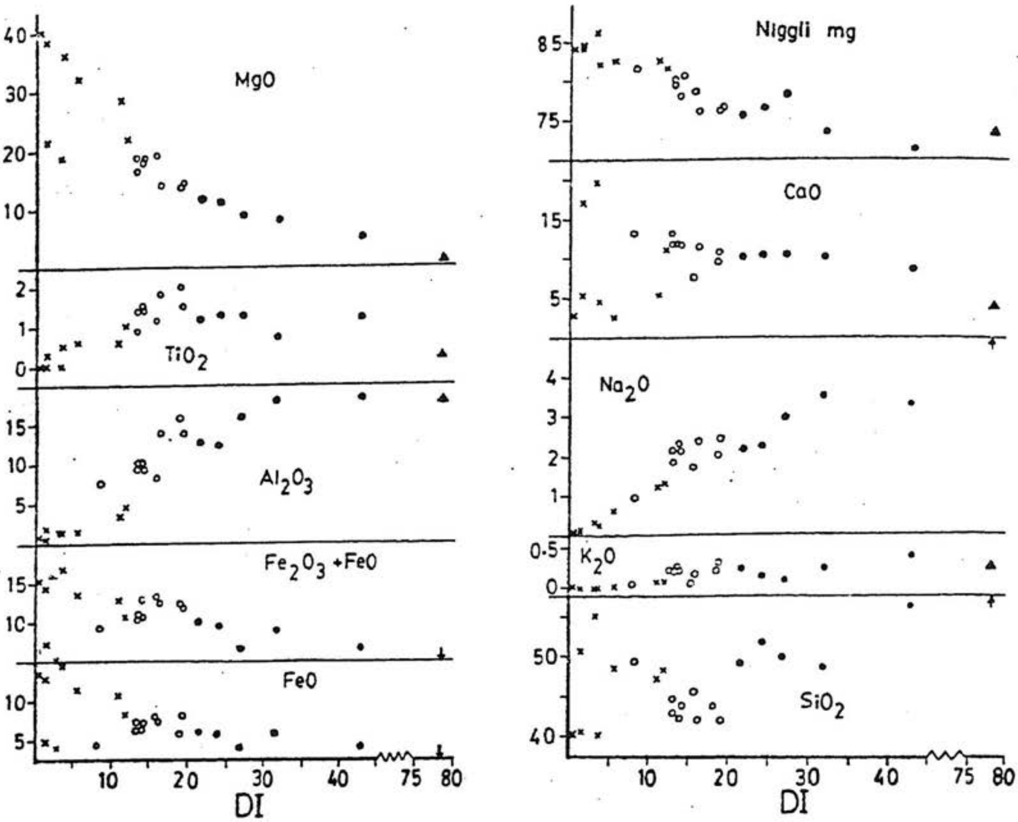


Fig. 2. Plots of oxide percentages and Niggli mg against differentiation index of Thornton & Tuttle (1960). Crosses: ultramafic rocks; circles: hornblendites; dots: metagabbros; triangle: plagiogranite.

the magma during early crystallization; Cr being preferentially accommodated in chromite and to a lesser extent in clinopyroxene, and Ni in olivine and to a lesser extent in ortho- and clinopyroxene, respectively. The strong depletion in these elements with differentiation suggests fractionation of olivine, spinel and pyroxenes at low pressure (8 kbar) (cf. Sivell & Rankin, 1983). Sr seems to have preferentially followed Ca in plagioclase and Zr shows a steady increase from ultramafic rocks to metagabbros through hornblendites.

Co and V show positive correlation with Fe^{2+} and Fe^{3+}, respectively, and with total iron. Since more iron was withdrawn in mafic phases as compared to liquid, Co and V (like iron in Fig.2) decrease with increasing DI. The values of the two elements in an olivine clinopyroxenite and that of V in a dunite fall expectedly below the variation trends of the remaining analyses. The variation of V against DI is very similar to that of TiO_2 (compare Fig. 2); in fact these two elements show a good positive correlation.

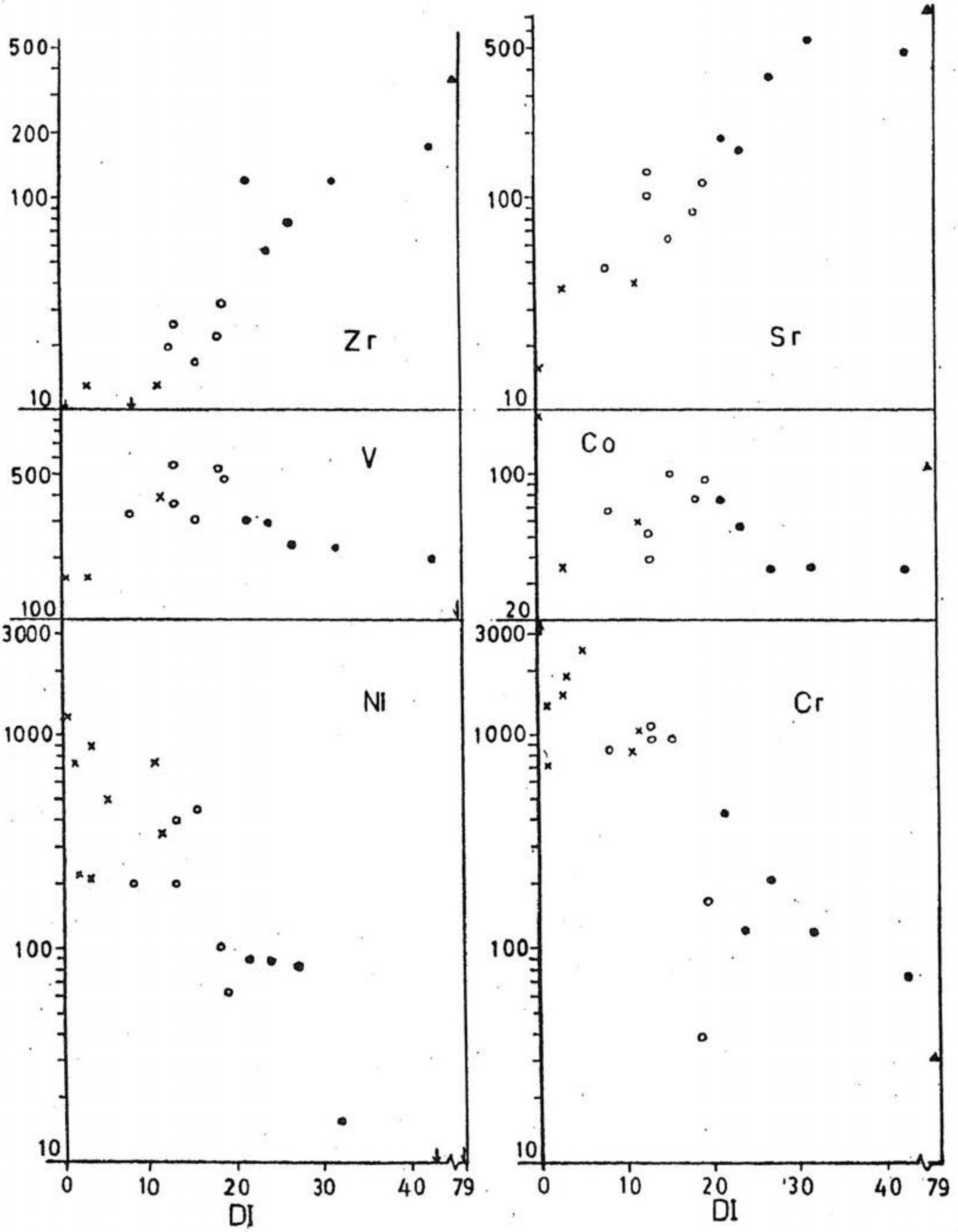


Fig. 3. Plots of selected trace elements (in ppm) against differentiation index of Thornton & Tuttle (1960). Symbols as in Fig. 2.

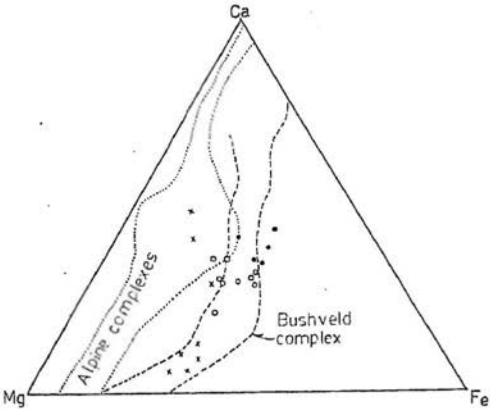
The observed variations in major, minor and trace elements are consistent with the hypothesis that the three groups of rocks are magmatically related. It would appear that olivine and chromite fractionation started earlier, joined and followed by ortho pyroxene, clinopyroxene and hornblende, leading to the formation of gabbros/norites and quartz diorites from the remaining melt. This led to successive formation of dunite, peridotites, olivine pyroxenites, pyroxenites, hornblende pyroxenites, pyroxene hornblendites, hornblendites, plagioclase hornblendites, (hornblende) gabbros/norites and pyroxene-quartz diorites. The plagiogranite might represent the residual liquid.

Tectonic environments of the rocks

Several diagrams were used to investigate the magmatic affinity of the rocks and tectonic environments prevailing during their formation. On MgO-Al_2O_3-CaO diagram of Coleman (1977) the hornblendites neither plot in the field of ophiolite ultramafic rocks nor in that of ophiolite gabbros. Six of the nine analyses fall within, or close to, the komatiite field. However, on other grounds the hornblendites cannot be considered komatiitic. Komatiites of similar SiO_2 contents generally contain higher MgO, Ni, Cr, and lower alkalis, TiO_2,Al_2O_3,CaO, Sr, V and Ba (cf. Nesbitt et al., 1979; Arndt & Nisbet, 1982). Besides, the immediately surrounding lithologies of the complex are not komatiitic.

The Mg-Fe-Ca plots of the analyses are presented in Fig. 4, together with fields for "alpine" rocks and Bushveld complex after Weedon (1970). The analyses are confined to the Bushveld field but the two pryoxenites plot in the alpine field because of their high CaO contents. On the Mg- (Fe^{2+} + Fe^{3+})-alk diagram of Weedon (1970), 18 of the Tora Tigga rocks are either confined to the Bushveld field or they plot in the overlapping area with the alpine complexes. On MgO-FeO^*-alkalis diagram (Fig. 5), the ultramafic rocks and six hornblendites plot in the field of ophiolite mafic-ultramafic cumulates (cf. Coleman, 1977) (This field, however, also incorporates many cumulate mafic-ultramafic rocks formed in other environments). But the plots of three hornblendites and the metagabbros outside this field are more significant and suggest that they may be non-ophiolitic. It can be seen that most of the analyses are confined to the field of mafic-ultramafic cumulate rocks from island arcs (Beard, 1986). The gabbros and hornblendites have calc-alkaline rather than tholeiitic character on this figure.

Skinner et al. (1978) compared the Bay of Islands ophiolite complex with Stillwater complex and found that the former has more Ni and higher Ni/Cr ratios. The six hornblendites (for which trace elements were determined), metagabbros and five ultramafic rocks plot in the field of Stillwater complex on Ni vs Cr diagram (Fig. 6). Only three ultramafic rocks plot above the Stillwater field but below that of the Bay of Islands ophiolite complex. Malpas & Stevens (1977) also found that alpine peridotites have higher NiO content (0.18) than layered rocks. On their NiO vs. Cr_2O_3 plot, all of the tora Tigga rocks plot in the area of the Stillwater layered intrusion.



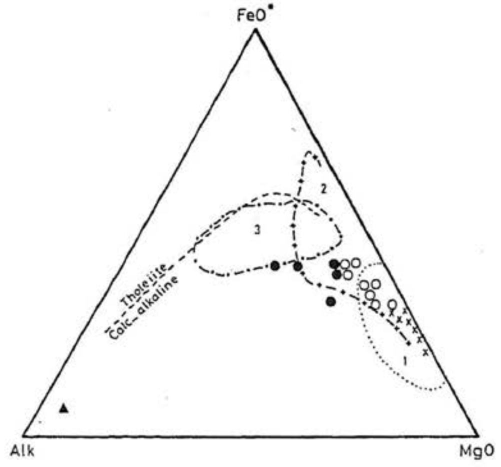


Fig. 4. (Left) Fe-Ca-Mg relations of the analyses, with fields for alpine-type rocks (dotted) and Bushveld complex (dashes) after Weedon (1970). Symbols as in Fig. 2.

Fig. 5. (Right) MFA plots of the analyses. Area 1 is for mafic-ultramafic cumulates in ophiolites (Coleman, 1977), area 2 for island arc cumulates and area 3 for island arc non-cumulates (Beard, 1986). Tholeiite-calc-alkaline boundary after Barker and Arth, 1976. Svmbols as in Fig. 2.

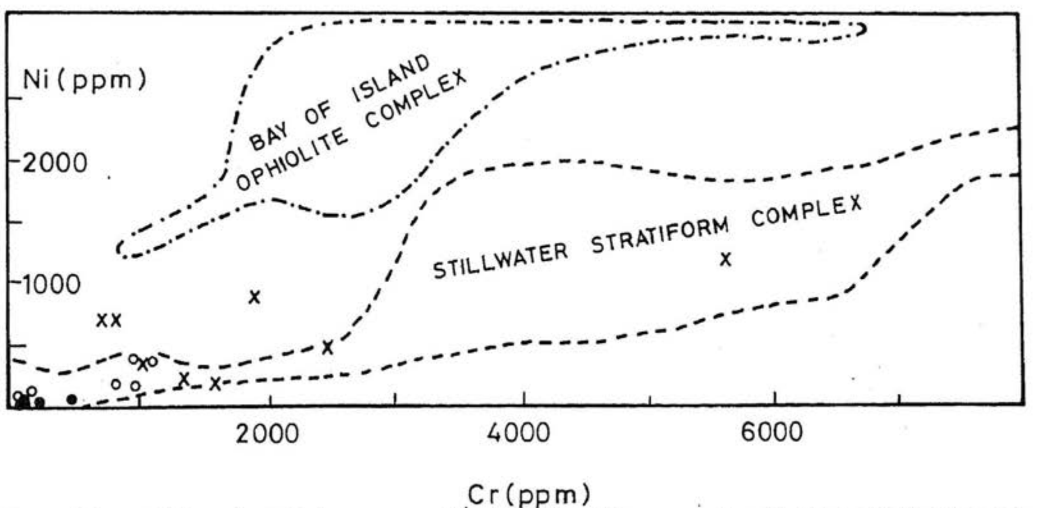


Fig. 6. Plots of Ni against Cr for some rocks of the Tora Tigga complex. Fields for Stillwater and Bay of Islands complexes are adapted form Skinner et al., 1978. Symbols as in Fig. 2.

Several major, minor and trace element diagrams have been proposed to discriminate between basic rocks of different tectonic environments. Although these are based on volcanic examples and their applicability to the Tora Tigga metagabbros is questionable, in the absence of other means they are relied upon here. The high degree of correlation and systematic variations, especially in ferromagnesian elements, suggest that their abundances have changed little during metamorphism (cf. Cann, 1969; Sivell & Rankin, 1983). The metagabbros in most cases plot in the field of island arc rocks on Ti vs Zr, Cr, V; Ni vs Y; Ti/Cr vs Ni; Zr/Y vs Zr; and Cr vs Ce/Sr diagrams (cf. Pearce & Cann, 1973; Pearce & Norry, 1979; Pearce, 1982).

Seven representative diagrams for the classification of the metagabbros are shown in Fig. 7 (for details see caption). In all, the rocks show island arc affinities. Only one analysis in Fig. 7C plots in MORB field because of its abnormally high Y content. Figs. 7B, F, and G suggest that the metagabbros have calc-alkaline affinity, but with a little lower Fe*/Mg ratios than typical calc-alkaline volcanic rocks. The K contents of the metagabbros are also significantly lower than typical calc-alkaline basalts but this may either be due to the cumulate nature of the metagabbros or explusion of K during metamorphic events (Sighinolfi & Gorgoni, 1978; Sheraton, 1984). the potassium content of the metagabbros, however, matches well with that of the hornblende-rich gabbros of the Rogue River island arc complex in southern Oregon (Garcia, 1982).

The multi-element spidergram of three selected hornblendites and a metagabbro is note-worthy (Fig. 8). The mantle-normalised values are characterised by distinct Nb trough, positive anomalies for Ba, P, and high K/Rb. The metagabbro also shows a peak for Sr. (Because of mineralogical control the hornblendites, on the other hand, display negative Sr and positive Ti anomalies). These features, notably Nb troughts, are commonly displayed by subduction-related magmas and quite similar to those of the Chilas complex (Khan et al., 1989). Comparison of data (Fig. 8B) for low-K tholeiite (Holm, 1985) and the active lavas from Mariana (Hole et al., 1984) suggests that the Tora Tigga rocks have originated in an Island arc setup.

DISCUSSION

Systematic variation in chemistry from ultramafic rocks to metagabbros through hornblendites suggests that the Tora Tigga rocks are derived from a basic magma by fractional crystallization. There are strong reasons to think that fractionation was controlled by early separation of olivine+minor chromite, followed by two pyroxenes, and hornblende. This led to an enrichment of SiO_2, Al_2O_3, alkalis, Sr, Zr, and depletion of MgO, FeO*, Cr, Ni, V and Co in the melt that produced gabbroic rocks.

Textural details suggest a replacement or metamorphic origin for much of the hornblende in the complex. The Na+K vs Al^{iv} plots of the hornblende in five hornblen

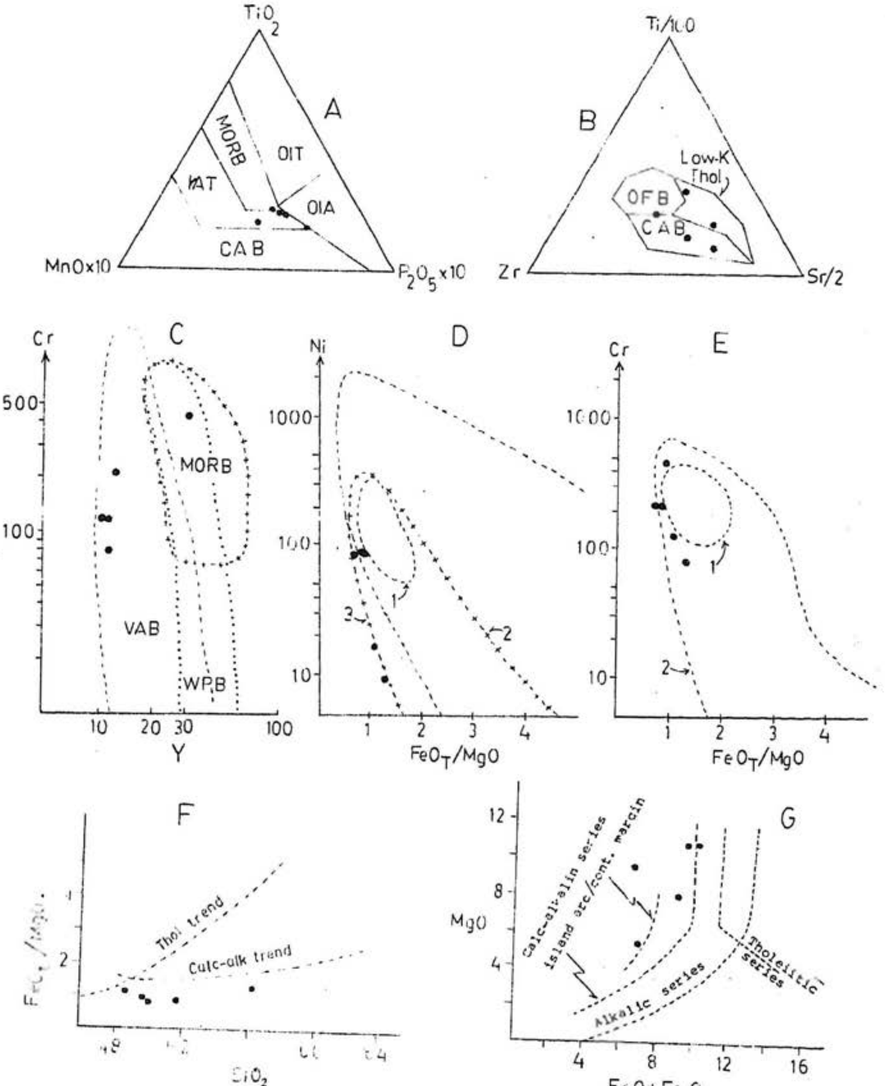


Fig. 7. Comparison of the Tora Tigga metagabbros with rocks of known tectonic environments. A is after Mullen (1984), B after Pearce and Cann (1973), C after Pearce (1982), D and E after Miyashiro and Shido (1973), F after Garcia (1982) and G after Yoder (1969). CAB: calc-alkaline basalts, IAT: island-arc tholeiites, MORB: mid-oceanic ridge basalts, OFB:ocean-floor basalts, OIA: oceanic-island alkali basalts, OIT: oceanic-island tholeiites, 1: abyssal tholeiites, 2: island arc and continental margin volcanic and plutonic rocks, 3: rocks of stable continents and oceanic islands. The tholeiitic trend in Fig. 7 F is based on Oman ophiolite and the calc-alkaline trend on Quaternary Mt. Hood volcanics. Symbols as in Fig. 2.

dites fall in the field of metamorphic rather than igneous huornblendes from the St. Anthony complex (Jamieson, 1981). It cannot, however, be ascertained that the hornblendites are metasomatic and chemically changed, because they show systematic chemical variations with the rest of the rocks. This discrepancy between textural details and chemistry can be explained in a number of ways: (1) The hornblendites and much of hornblende in the remaining rocks are igneous but recrystallized during metamorphism, (2) There are igneous as well as metamorphic hornblendes, and the chemical data are not sufficient to distinguish the origin of the amphiboles in the complex.

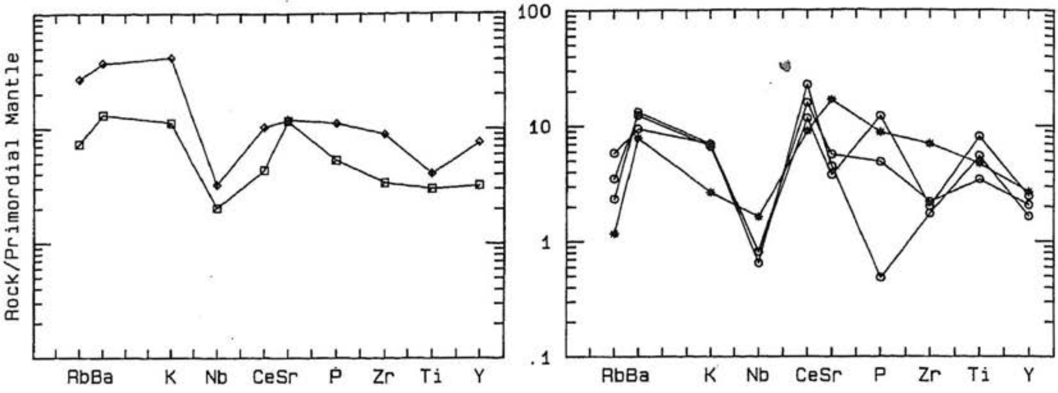


Fig. 8. Multi-element plots normalised to primordial mantle (cf. Wood, 1979). The data for metagabbro (tt 14, asteric) and three hornblendites (tt 17,96,139, circles) from Tora Tigga (right) compares well with those of islannd arcs (left). Low-K tholeiite (square) after Holm (1985) and present day lavas (rhomb) from Mariana island after Hole et al. (1984).

We are unaware of the occurrence of such large bodies of hornblendites in "alpine" or stratiform complexes. However, large bodies of hornbldnee-rich gabbros and diorites occur elsewhere in the world. In the plutonic blocks from Lesser Antilles island arc (Arculus & Wills, 1980), modal hornblende reaches up to 70%. Garcia (1982) has reported a gabbro forming the basal part of a Cretaceous island arc in SW Oregon. This body is essentially composed of hornblende (about 50%) and calcic plagioclase, and is therefore similar to some of the plagioclase hornblendites of Tora Tigga. Hornblende, generally, is abundant and hornblendites are more common in the Alaskan-type complexes (Taylor, 1967) confined to gabbroic terrains. However, a complete lack of hornblende-magnetite clinopyroxenites, abundance of orthopyroxene, and the low Al_2O_3 content of the clinopyroxene (deduced from the whole rock analyses of clinopyroxene-rich rocks) do not typify the Tora Tigga complex as Alaskan-type.

The Duke Island Alaskan-type ultramafic complex contains variable amounts of post-cumulus replacive-type hornblende in ultramafic rocks, as well as marginal hornblendites, and hornblende-Ca-rich plagioclase pegmatite swarms. The hornblendites are variable in grain-size and pegmatitic varieties may cut the others (Irvine, 1974). In these respects the Duke Island and Tora Tigga complexes have a close resemblance. Irvine suggested that the hornblende was derived from an intercumulus magma that was rich in water and different from the settled minerals. The hornblendites and hornblende pegmatites were produced from this liquid. Such a situation may also have existed in Tora Tigga, the final manifestation of which may be the plagiogranite dykes that are invariably associated with the hornblendites. The production of so much hornblende and hornblendites from intercumulus liquid would require a larger quantity of ultramafic rocks in Tora Tigga. It is possible that ultramafic rocks are more abundant at depth, since the base of the complex is not exposed.

In conclusion, it is likely that the hornblendites and at least some hornblende in the remaining rocks are of igneous origin. The quantity of hornblende, however, may have increased during metamorphism. Much of the hgornblende in metagabbros, for example, clearly has grown at the expense of pyroxene during amphibolite facies metamorphism.

The Tora Tigga ultramafic rocks have closer similarities with stratiform than tectonized ultramafic rocks, and the metagabbros have island arc-type chemical characteristics. Their rapidly decreasing FeO*, TiO, V and increasing SiO_2, Al_2O_3 contents with differentiation suggest that they may be derived from a calc-alkaline magma. The Chilas complex, forming the back-bone of the Kohistan arc, is also considered calc-alkaline (Jan, 1980; Bard, 1983). In a number of places in the amphibolite belt, there are partially amphibolitized bodies of plutonic rocks identical to those of the Chilas complex; some of these are located not far from Tora Tigga. Thus, and as already suggested on the basis of petrographic similarities, the Tora Tigga complex may be related to the Chilas complex of Khan et al. (1989).

Luhr & Carmichael (1985) regard high-alumina basalts too depleted in MgO, Ni, and Cr to have been generated by partial melting of mantle, and that significant fractionation of olivine, augite, plagioclase and spinel has probably played a role in their evolution from primitive basalt magma. The Tora Tigga and (parts of) Chilas complexes may represent such an example of fractionated minerals in the base of an island arc.

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REFERENCES

Andrews-Speed, C.P. & Brookfield, M.E., 1982. Middle Palaeozoic to Cenozoic geology and tectonic evolution of northwestern Himalaya. Tectonophysics 8, 253-275.

Arculus, R.J. & Wills, K.J.A., 1980. The petrology of plutonic blocks and inclusions from the Lesser Antilles island arc. J. Petrology 21, 743-799.

Arndt, N.T. & Nisbet, E.G., 1982. What is komatiite? In: Komatiites (N.T. Arndt & E.G. Nisbet, eds). Allen & Unwin, London, 19-27.

Banaras, M. & Ghani, A., 1982. Petrography of the Tora Tigga complex, Munda area, Dir district. M.Sc. Thesis, Peshawar University.

Bard, J.P., 1983. Metamorphic evolution of an obducted island arc: Example of the Kohistan sequence (Pakistan) in the Himalayan collided range. Geol. Bull. Univ. Peshawar 16, 105-184.

Barker, E. & Arth, JJ.G., 1976. Generation of trondhjemitic- tonalitic liquids and Archean bimodal trondhjemite-basalt suites. Geology 4, 596-600.

Beard, J.S., 1986. Characteristic mineralogy of arc-related cumulate gabbros; implications for the tectonic seting of gabbroic plutons and for andesite genesis. Geology 14, 848-851.

Butt, K.A., Chaudhry, M.N. & Ashraf, M., 1980. An interpretation of petrotectonic assemblage west of W. Himalayan syntaxis in Dir District and adjoining areas, northern Pakistan. Geol. Bull. Univ. Peshawar 13, 79-86.

Cann, J.R., 1969. Spilites from the Carlsberg Ridge, Indian Ocean. J. Petrology 10, 1-9.

Chaudhry, M.N., Kausar, A.B. & Lodhi, S.A.K., 1974. Geology of Timurgara-Lal Qila area, Dir District, N.W.F.P. Geol. Bull. Punjab Univ. 11, 53-73.

Coleman, R.G., 1977. Ophiolites-Ancient Oceanic Lithosphere. Springer, Berlin.

Coleman, R.G. & Keith, T.E., 1971. A chemical study of serpentinization, Burro Mountain, California. Jour. Petrology 12, 311-328.

Coward, M.P., Jan, M.Q., Rex, D., Tarney, J., Thirlwall, M. & Windley, B.F., 1982. Geotectonic frame-work of the Himalaya of N. Pakistan. Jour. Geol. Soc. Lond. 139, 299-308.

Coward, M.P., Windley, B.F., Broughton, R., Luff, I.W., Petterson, M., Pudsey, C., Rex, D. & Khan, M.A., 1986. Collision tectonics in the N.W. Himalayas. In: Collision Tectonics (M.P. Coward & A.C. Ries, eds.). Geol. Soc. Lond. Sp. Publ. 19, 203-219.

Garcia, M.Q., 1982. Petrology of the Rogue River island arc complex, southwestern Oregon. Amer. Jour. Sci. 182, 783-807.

Green, D.H., 1964. The petrogenesis of the high-temperature peridotite intrusion in the Lizard area, Comwall. Jour. Petrology 5, 134-188.

Himmelberg, G.R. & Loney, R.A., 1973. Petrology of the Vulcan Peak alpine-type peridotite, southwestern Oregon. Geol. Soc. Amer. Bull. 84, 1585-1600.

Hole, M.J., Saunders, A.D., Mariner, G.F. & Tarney, J., 1984. Subduction of pelagic sediment; Implications for the origin of Ce-anomalous basalts from the Mariana Islands. Jour. Geol. Soc. Lond. 141, 453-472.

Holm, P.E., 1985. The geochemical fingerprints of different tectonomagmatic environments using hygromagmatophile element abundances of tholeiitic basalts and basaltic andesites. Chem. Geol. 51, 303-323.

Honegger, K., Dietrich, V., Frank, W., Gansser, A., Thoni, M. & Trommsdorff, V., 1982. Magmatism and metamorphism in the Ladakh Himalayas (the Indus-Tsangpo suture zone). Earth Planet. Sci. Letters 60, 253-292.

Irvine, T.N., 1974. Petrology of the Duke Island Ultramafic Complex, Southeastern Alaska. Geol. Soc. Amer. Mem. 138.

Irvine, T.N. & Baragar, W.R.A., 1971. A guide to the classification of the common volcanic rocks. Can. J. Earth Sci. 8, 523-549.

Jackson, E.D., 1961. Primary textures and mineral association in the ultramafic zone of the stillwater complex, Montana. U.S.G.S. Prof. Pap. 358, 1-106.

Jackson, E.D. & Thayer, T.P., 1972. Some criteria for distinguishing stratiform, concentric and alpine peridotite-gabbro complexes. 24th Int. Geol. Cong. 2, 289-296.

Jamieson, R.A., 1981. Metamorphism during ophiolite emplacement, the petrology of St. Anthony Complex. Jour. Petrology 22, 397- 449.

Jan, M.Q., 1977. The mineralogy, geochemistry and petrology of Swat Kohistan, NW Pakistan. Ph.D. Thesis, Univ. London.

Jan, M.Q., 1980. Petrology of the obducted mafic and ultramafic metamorphites from the southern part of the Kohistan island arc sequence. Geol. Bull. Univ. Peshawar 13, 95-108.

Jan, M.Q., Banaras, M., Ghani, A. & Asif, M., 1983. The Tora Tigga ultramafic complex, southern Dir district. Geol. Bull. Univ. Peshawar 16, 11-29.

Jan, M.Q. & Howie, R.A., 1980. Ortho- and clinopyroxenes from granulites of Swat Kohistan, northern Pakistan. Min. Mag. 43, 715-728.

Jan, M.Q. & Howie, R.A., 1981. The mineralogy and geochemistry of the metamorphosed basic and ultrabasic rocks of the Jijal Complex, Kohistan, N.W. Pakistan. Jour. Petrology 22, 85-126.

Jan, M.Q., Khattak, M.U.K., Parvez, M.K. & Windley, B.F., 1984. The Chilas stratiform complex: Field and mineralogical aspects. Geol. Bull. Univ. Peshawar 17, 153-169.

Kakar, S.K., Badshah, M.S. & Khan, J., 1971. The Geology of the Jandul Valley, western Dir. Geol. Bull. Univ. Peshawar 6, 54-73.

Khan, M.A., Jan, M.Q., Windley, B.F., Tarney, J. & Thirlwall, M.F., 1989. The Chilas mafic-ultramafic igneous complex; The root of the Kohistan island arc in the Himalaya of northern Pakistan. Geol. Soc. Am. Spec. Paper 232, 75-94.

Kuno, H., 1959. Origin of Cenozoic petrographic provinces of Japan and surrounding areas. Bull. Volcanol., Ser II, 533-563.

Leake, B.E., 1964. The chemical distrinction between ortho- and para-amphibolites. Jour. Petrology 15, 238-254.

Leake, B.E., 1978. Nomenclature of amphiboles. Min. Mag. 42, 533-563.

Loney, R.A., Himmelberg, G.R. & Coleman, R.G., 1971. Structure and petrology of alpine peridotite at Burro Mountain, California, USA. Jour. Petrology 12, 245-309.

Luhr, J.F. & Carmichael, I.S.E., 1985. Jorullo Volcano, Michoacan, Mexico (1759-1774): The earliest stages of fractionation in calc-alkaline magmas. Contrib. Mineral. Petrol. 90, 142-161.

Miyashiro, A., 1975. Classification, characteristics, and origin of ophilites. Jour. Geology 83, 249-281.

Miyashiro, A. & Shido, F., 1975. Tholeiitic and calc-alkaline series in relation to the behaviour of titanium, vanadium, chromium and nickel. Amer. Jour. Sci. 275, 265-278.

Malpas, J. & Stevens, R.K., 1977. The origin and emplacement of the ophiolite suite with examples from western Newfoundland. Geotectonics 11, 453-466.

Moore, E.M., 1973. Geotectonic significance of ultramafic rocks. Earth. Sci. Rev. 9, 241-258.

Mullen, E.D., 1983. $MnO/TiO_2/P_2O_5$: a minor element discriminant for basaltic rocks of oceanic environment and its implication for petrogenesis. Earth Planet. Sci. Letters 62, 53-62.

Nesbitt, R.W., Sun, S.S. & Purvis, A.C., 1979. Komatiites: Geochemistry and genesis. Can. Mineral. 17, 165-186.

Pearce, J.A., 1982. Trace element characteristics of lavas from destructive plate boundaries. In: Andesites (R.S. Thorpe. ed.). Wiley, New York, 525-548.

Pearce, J.A. & Cann, J.R., 1973. Tectonic setting of basic volcanic rocks determined using trace element analyses. Earth Planet Sci. Letters 19, 290-300.

Pearce, J.A., Harris, N.B.W. & Tindle, A.G., 1984. Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. Jour. Petrology 25, 956-983.

Pearce, J.A. & Norry, M.J., 1979. Petrogenetic implications of Ti, Zr, Y and Nb variations in volcanic rocks. Contrib. Mineral. Petrol. 69, 33-47.

Petterson, M.G. & Windley, B.F., 1986. Petrological and geochemical evolution of the Kohistan arc-batholith, Gilgit, N. Pakistan. Geol. Bull. Univ. Peshawar 19, 121-149.

Sheraton, J.W., 1984. Chemical changes associated with high-grade metamorphism of mafic rocks in the East Antarctic sheld. Chem. Geol. 47, 135-157.

Sighinolfi, G.P. & Gorgoni, C., 1978. Chemical evolution of high-grade metamorphic rocks--anatexis and remotion of material from granulite terains. Chem. Geol. 22, 157-176.

Sivell, W. & Rankin, P., 1983. Arc-tholeiite and ultramafic cumulate, Brook Street volcanics, west D'Urville Island, New Zealand. N.Z.J. Geol. Geophys. 26, 239-257.

Skinner, W.R., Bowes, D.R. & Skinner, D.L., 1978. Petrochemistry of chromite-bearing ultramafic complexes, Red Lodge, Montana, USA. In: Mineralization in Metamorphic Terranes (W.J. Verwood, ed.). Van Schaik, Pretoria, 509-527.

Spry, A., 1969. Metamorphic Textures. Pergamon Press, Oxford.

Tahirkheli, A.Z., 1983. Geochemistry of the Tora Tigga complex, southern Dir district. M.Sc. Thesis, Univ. Peshawar.

Tahirkheli, R.A.K. & Jan, M.Q. (eds.), 1979. Geology of Kohistan, Karakoram Himalaya, Northern Pakistan. Geol. Bull. Univ. Peshawar (Spec. Issue) 11.

Tahirkheli, R.A.K., Mattauer, M., Proust, F. & Tapponnier, P., 1979. The India-Eurasia suture zone in northern Pakistan, some new data and interpretation at plate scale. In: Geodynamics of Pakistan (A. Farah & K.A. DeJong, eds.). Geol. Surv. Pak. Quetta, 125-130.

Taylor, H.P., 1967. The zoned ultramafic complexes of southeastern Aklaska. In: Ultramafic and Related Rocks (P.J. Wyllie, ed.) Wiley, New York, 97-121.

Thornton, C.P. & Tuttle, O.F., 1960. Chemistry of igneous rocks, 1. Differentiation index. Amer. Jour. Sci. 258, 664-684.

Tuttle, O.F. & Bowen, N.L., 1958. Origin of granite in the light of experimental studies in the system Na Al Si_3O_8-K Al Si_3O_8-SiO_2-H_2O. Geol. Soc. Amer. Mem. 74.

Vernon, R., 1968. Microstructures of high grade metamorphic rocks at Broken Hill, Australia. Jour. Petrology 9, 1-22.

Wager, L.R. & Deer, W.A., 1939. Geological investigation in East Greenland: pt.III, The petrology of the Skaergaard intrusion, Kangardlugsauq, East Greenland. Medd. Gronland 105, 1-352.

Weedon, D.S., 1970. The ultrabasic-basic igneous rocks of the Huntley region. Scot. Jour. Geol. 16, 26-40.

Wood, D.A., 1979. A variably veined suboceanic upper mentle -- Genetic significance for mid-ocean ridge basalts from geochemical evidence. Geology 7, 499-503.

Wright, J.B., 1969. A simple alkalinity ratio and its application to questions of non-orogenic granite genesis. Geol. Mag. 106, 370-384.

Yoder, H.S., 1969. Calc-alkaline andesites: experimental data bearing on the origin of their assumed characteristics. In: Proceedings of the Andesite Conference (A.R. McBirney, ed.) Oregon Dept. Geol. Mineral Industry Bull. 65, 77-89.