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Ortho- and clinopyroxenes from the pyroxene granulites of Swat Kohistan, northern Pakistan

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SUMMARY. This paper presents wet chemical analyses of forty-one pyroxenes (thirty-nine of them new) from pyroxene granulites and a few neighbouring rocks from Swat Kohistan and the adjoining Indus Valley. The granulites, considered to be derived from plutonic norites of an island arc tholeiitic nature, constitute one of the most extensive belts of its kind in the world. The pyroxenes are not unusual in any respect but they bear a closer resemblance to metamorphic than to igneous pyroxenes crystallized from deep-seated tholeiitic magmas. The distribution of Mg, Fe, and Mn (average $K_D^{\text{Mg-Fe}} = 0.57$) and the tie-line intersections on the Wo-En join (generally from $Wo_{80.5}$ to $Wo_{76.5}$) between the coexisting eighteen pyroxene pairs from the granulites are discussed. Based on eight different methods of geothermometry and other considerations, it is concluded that the pyroxene granulites were metamorphosed at around 800 °C and 7 to 8 kbar.

AN exceptionally large belt of pyroxene granulites, possibly the most extensive in the world, occurs in Swat Kohistan and the adjacent districts of northern Pakistan. The rocks stretch for about 200 km in an approximately 35 km wide belt between Nanga Parbat and northern Swat, tapering westwards into Dir (see map of Swat Kohistan in Jan and Kempe, 1973). In the south the granulites are bordered by amphibolites, the latter probably representing an oceanic crust and having a tectonic contact with metasedimentary and plutonic (mainly granitic) rocks of the Indian plate to the south (Jan, 1977*a*). To the north the granulites are bordered by a belt of quartz diorite plutons which were probably generated by partial melting during the collision of the Indian-Eurasian plates and subsequent Himalayan orogenic episodes (Jan, 1977*b*).

The granulites are medium- to coarse-grained gneissose rocks with a general E-W to NE-SW foliation dipping to the north. They were previously considered to be slightly or non-metamorphosed norites (Jan and others, 1970, 1971, 1973; Chaudhry and Chaudhry, 1974; Desio, 1974; Shams, 1975). However, the petrography and mineralogy of the rocks are strongly suggestive of granulite facies metamorphism of plutonic basic to intermediate (mainly noritic) rocks derived from a tholeiitic magma of island arc/continental margin affinity. The noritic rocks, being the plutonic equivalents of the tholeiitic (calc-alkaline) volcanic series, may have originally formed in an island arc in the ancient Tethyan sea (Jan and Kempe, 1973, Tahirkheli *et al.*, 1979), or they may have formed during the Archaean as part of earlier major crust-building activity (Jan, 1977*b*).

The granulites, mostly represented by leucocratic facies, are essentially composed of plagioclase (An_{45-55}, generally antiperthitic) and ortho- and clinopyroxenes, with minor amounts of opaque minerals, hornblende, biotite, and apatite. Quartz is present in most rocks, its amount increasing to an essential component in the intermediate members which also contain perthitic K-feldspar. The biotite is replacive and appears to be secondary but it cannot be ascertained whether or not all the amphibole is also secondary. Olivine is absent even in the pyroxenite members and in olivine-normative rocks; so also is rutile, found only rarely as inclusions in the plagioclase. Garnet occurs in only one place, to the north of Seo, along the Indus. In this area of granulites garnet-bearing pegmatites are also found. Throughout the region, especially in Swat, the granulites display banding; most of the bands are noritic but some are pyroxenitic or anorthositic. Some appear to be relict igneous layers distorted and modified by tectonism and metamorphism, whilst others seem to be a product of metamorphic differentiation. Retrograde alteration of the granulites to amphibolite is common, especially at their southern margin in the Madyan-Fatehpur area.

This paper presents a detailed discussion of forty-one ortho- and clinopyroxene analyses (thirty-nine of them new) from the area. Thirty-five of the analyses are from the pyroxene granulites, two from a hornblende websterite granulite (US 24) in amphibolites, three from quartz diorites, and one from a calc-silicate rock, thirty-eight of them being in coexisting pairs. The granulite pyroxenes have been compared with other metamorphic examples, and igneous pyroxenes of tholeiitic plutons, and the $K_D^{\text{Mg-Fe,Mn}}$ and tie-line intersections have been discussed along with T–P estimates for the rocks.

Orthopyroxenes. Orthopyroxene is the most abundant ferromagnesian mineral of the granulites. Chemical analyses of twenty orthopyroxenes are presented in Table I; analysis SH 16 is taken from Jan and Kempe (1973) and included here for the sake of completeness. US 24 is from a hornblende websterite (granulite) mass in the amphibolites and 448 from a quartz diorite, respectively, to the south and north of the granulite belt. Analysis 647 is also from a dioritic rock to the north. The status of this rock is not clearly understood but it may be a retrograde intermediate granulite containing much secondary hornblende enclosing hypersthene relics. The remaining seventeen analyses are from the pyroxene granulites.

The analyses, together with an optically determined value from a pyroxenite member (SK 340) of the granulites, have been plotted in the pyroxene diagram (fig. 1). The composition ranges from En_{64} to En_{51} but those from the pyroxenites are more magnesian (En_{75} and En_{78}). The En content of their orthopyroxenes depends strongly on the Fe:Mg ratio of their host rocks. The density, refractive indices and, in general, depth of colour increase with decrease in En content. A plot of En *vs* density produces a trend with slightly increased gradient compared with that in Deer *et al.* (1978, p. 109); the two trends intersect at En_{59}.

Like other orthopyroxenes of the granulite facies rocks (Howie, 1965), the orthopyroxenes of Swat granulites are strongly pleochroic with α pink or reddish pink, β yellowish, and γ green. The strength of pleochroism has been attributed to one or more of the ions Fe^{3+}, Fe^{2+}, Ti, Mn, Cr, Ni, and Al by various workers, to diopside lamellae by Parras (1958), and to lattice shortening/distortion essentially in response to a greater Al substitution (Howie, 1963, 1965; Davidson and Mathison, 1974). The phenomenon of pleochroism is complex and beyond the scope of this paper. However, the abundance of any one single element cannot satisfactorily explain it; it seems to be a combined effect of the various elements in the octahedral site, coupled with the amount of Al^{iv}. Burns (1970) has explained orthopyroxene pleochroism in terms of crystal field theory and crystal chemistry.

Some degree of exsolution is displayed by the ortho- as well as the clinopyroxenes of Kohistan; the exsolved material forms irregular blebs of varying size and shape distributed along planes parallel to (100) of the host grains. In addition very thin lamellae are also found in a few cases. Exsolved clinopyroxene in orthopyroxene of the stratiform complexes, on the other hand, is in more continuous blebs and uniformly thick lamellae, and is more frequent and abundant (Poldervaart and Hess, 1951; Brown, 1957; Wager and Brown, 1968). In some of the rocks the pyroxenes do not display exsolution, but in these the phenomenon may be on a submicroscopic scale, analogous to cryptoperthites. In a few cases the exsolved blebs are concentrated in the strained parts (i.e. along axes of deformation) of the host pyroxene. This leads to the suggestion that exsolution took place either during metamorphism, or that previously exsolved lamellae were modified during deformation. The common development of exsolution phenomenon in the Broken Hill metamorphic pyroxenes has been regarded by Binns (1965) as suggesting that ordering in the Ca-rich and Ca-poor pyroxenes has been thorough.

The pyroxene analyses in Table I suggest that with increase in Fs content there is an increase in Mn, decrease in Si and, less obviously, in Al. There also seems to be a general increase in CaO with increasing Fe, a relationship previously suggested by the work of Howie and Smith (1966), Davidson (1968), and Scharbert and Kuart (1974); Dobretsov *et al.* (1972) have found a positive correlation between Al_2O_3 and Fe_2O_3. This relationship is not very obvious in the Kohistan orthopyroxenes (including five analyses from the Jijal complex; Jan and Howie, in preparation); however, the average Fe_2O_3 content (1.7 %) of the orthopyroxenes with less than 2 % Al_2O_3 is slightly lower than that (1.8 %) in those with more than 2 % Al_2O_3. Among the trace elements, Cr and Ni generally fall with decrease in Mg. A comparison of the orthopyroxene analyses with those of the coexisting clinopyroxene (Table II) shows that the former, in general, are richer in SiO_2, MnO, Co, Cu, Ni, Zn, and poorer in Al_2O_3, Na_2O, Fe_2O_3, TiO_2, P, Cr, Li, V, and in Mg/Fe^{2+} ratio. These observations are in agreement with those summarized in Deer *et al.* (1978).

The amount of Al^{iv} is invariably higher than Al^{vi}, as also noted by Howie (1955) and Dobretsov *et al.* (1972). The Mn, Cr, and Ni contents fall within the field of orthopyroxenes investigated by Howie

TABLE I **Analyses of the orthopyroxenes***

	US 24	SK 322	SI 192	SH 16	SK 647	SK 600	US 9	SK 422	SK 528	SK 592	SK 448	SI 188a	SK 615	SI 188b	SI 197	US 15	SI 182	US 14	SI 193	SI 221
SiO_2	53.60	50.93	52.05	50.80	51.92	51.01	51.27	51.09	50.85	51.73	51.54	51.13	51.64	51.33	51.35	51.02	50.44	51.10	50.81	50.90
TiO_2	0.21	0.38	0.26	0.15	0.26	0.34	0.33	0.26	0.25	0.31	0.58	0.28	0.17	0.33	0.22	0.21	0.31	0.33	0.27	0.34
Al_2O_3	2.42	3.12	1.67	2.33	1.36	2.13	2.47	1.94	2.89	1.88	2.12	2.03	1.64	2.35	1.62	1.95	2.67	1.61	1.33	1.28
Fe_2O_3	1.42	1.67	2.58	1.86	1.82	2.34	1.80	1.61	2.79	1.36	1.82	1.38	1.93	1.42	2.45	0.96	1.35	2.68	1.78	1.27
FeO	14.59	19.99	19.81	21.21	21.17	21.35	22.14	22.44	21.25	22.53	20.56	22.98	22.55	22.93	22.77	24.08	24.03	23.56	25.20	27.02
MnO	0.37	0.53	0.52	0.49	1.05	0.54	0.52	0.73	0.54	0.59	0.60	0.66	0.68	0.76	0.62	0.62	0.69	0.67	0.83	0.84
MgO	26.85	21.71	21.77	21.17	20.81	20.71	20.54	20.44	20.12	19.89	18.42	20.00	19.70	19.26	19.52	19.27	19.48	18.67	18.12	17.20
CaO	0.99	1.41	1.11	1.36	1.09	1.03	0.87	1.04	1.04	1.31	3.58	1.24	1.11	1.40	1.11	1.60	0.82	1.35	1.25	1.24
Na_2O	0.01	0.08	0.05	0.11	0.02	0.02	0.03	0.03	0.03	0.04	0.13	0.05	0.08	0.11	0.03	0.05	0.07	0.04	0.02	0.07
K_2O	0.00	0.00	0.01	0.01	0.00	0.00	0.01	0.00	0.02	0.04	0.06	0.02	0.01	0.01	0.01	0.03	0.01	0.01	0.02	0.04
H_2O+	-	-	-	0.45	-	-	-	-	-	-	0.25	-	-	-	-	0.17	-	-	-	-
P_2O_5	0.00	0.00	0.00	-	0.02	0.00	0.00	0.04	0.00	0.00	0.02	0.00	0.01	0.04	0.00	0.04	0.00	0.00	0.10	0.00
TOTAL	100.46	99.82	99.83	99.94	99.52	99.47	99.98	99.62	99.78	99.68	99.68	99.77	99.52	99.94	99.70	100.00	99.87	100.02	99.73	100.20
α	1.683	-	1.694	-	-	1.696	-	-	1.699	-	-	1.700	-	-	-	-	-	1.702	-	1.706
β	1.691	-	1.706	-	-	1.708	-	-	-	-	-	1.712	-	-	-	-	-	1.714	-	1.719
γ	1.695	-	1.710	-	-	1.713	-	-	1.714	-	-	1.716	-	-	-	-	-	1.718	-	1.723
D	3.403	3.457	3.482	-	3.496	3.493	3.501	3.496	3.495	3.475	3.415	-	3.506	3.513	3.520	3.542	3.550	3.549	3.562	3.576
NUMBERS OF IONS ON THE BASIS OF SIX OXYGENS																				
Si	1.926	1.899	1.939	1.895	1.953	1.922	1.923	1.930	1.910	1.949	1.946	1.932	1.952	1.937	1.943	1.937	1.912	1.939	1.947	1.953
Al	0.074	0.101	0.061	0.103	0.047	0.078	0.077	0.070	0.090	0.051	0.054	0.068	0.048	0.063	0.057	0.063	0.088	0.061	0.053	0.047
Al	0.029	0.036	0.012	-	0.013	0.017	0.032	0.016	0.038	0.033	0.040	0.022	0.025	0.041	0.015	0.024	0.031	0.011	0.007	0.011
Ti	0.006	0.011	0.007	0.004	0.007	0.010	0.009	0.007	0.007	0.009	0.016	0.008	0.005	0.009	0.006	0.006	0.009	0.009	0.008	0.010
Fe^{3+}	0.038	0.047	0.072	0.052	0.052	0.066	0.051	0.046	0.079	0.039	0.052	0.039	0.055	0.040	0.070	0.027	0.039	0.077	0.051	0.037
Fe^{2+}	0.438	0.623	0.617	0.662	0.666	0.673	0.695	0.709	0.668	0.710	0.649	0.726	0.713	0.723	0.721	0.765	0.762	0.748	0.808	0.867
Mn	0.011	0.017	0.016	0.016	0.034	0.017	0.017	0.023	0.017	0.019	0.019	0.021	0.022	0.024	0.020	0.020	0.022	0.022	0.027	0.027
Mg	1.438	1.206	1.209	1.177	1.167	1.163	1.148	1.151	1.127	1.117	1.037	1.126	1.110	1.083	1.101	1.090	1.101	1.056	1.035	0.984
Ca	0.038	0.056	0.044	0.054	0.044	0.042	0.035	0.042	0.042	0.053	0.144	0.050	0.045	0.057	0.045	0.065	0.035	0.055	0.051	0.051
Na	0.001	0.006	0.004	0.008	0.002	0.002	0.002	0.002	0.002	0.003	0.010	0.004	0.006	0.008	0.002	0.004	0.005	0.003	0.002	0.005
K	0.000	0.000	0.001	0.001	0.000	0.000	0.001	0.000	0.001	0.002	0.003	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.002
Z	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000
XY	1.999	2.002	1.982	1.974	1.985	1.990	1.990	1.996	1.981	1.985	1.970	1.997	1.982	1.986	1.981	2.002	2.003	1.982	1.990	1.994
mg†	74.7	63.7	63.2	61.7	60.8	60.60	60.07	59.66	59.60	59.26	59.02	58.89	58.42	57.91	57.58	57.31	57.22	55.49	53.88	51.38
Mg	73.2	61.9	61.7	60.0	59.5	59.3	59.0	58.4	58.3	57.7	54.5	57.4	57.1	56.2	56.3	55.4	56.3	54.0	52.5	50.00
Fe	24.9	35.2	36.0	37.2	38.3	38.6	39.2	39.5	39.5	39.6	37.9	40.0	40.6	40.9	41.4	41.3	42.0	43.2	44.9	47.4
Ca	1.9	2.9	2.3	2.8	2.2	2.1	1.8	2.1	2.2	2.7	7.6	2.6	2.3	2.9	2.3	3.3	1.7	2.8	2.6	2.6
PPM																				
Co	94	117	117	-	88	117	116	104	125	102	78	118	109	113	112	110	103	108	97	76
Cr	1590	103	114	400	22	105	155	28	81	81	363	37	24	12	83	56	84	69	23	29
Cu	24	111	37	-	56	44	52	31	43	14	76	166	21	175	57	104	17	32	182	56
Li	1	4	3	-	4	5	4	6	6	2	10	4	6	4	5	7	3	8	3	2
Ni	374	176	213	300	61	173	211	112	193	105	119	82	100	61	148	119	96	154	78	79
Pb	11	5	<5	-	<5	<5	<5	6	<5	<5	26	<5	<5	<5	<5	<5	<5	<5	<5	<5
V	98	207	141	300	50	164	176	128	201	144	144	173	105	135	159	152	182	158	128	100
Zn	209	-	364	-	472	390	396	486	401	337	-	466	443	482	440	461	356	480	531	428

* US 24: From hornblende websteritic granulite in amphibolites
SK 448: From quartz diorite
SK 647: From dioritic rock (probably retrograde granulite)

The remaining orthopyroxenes are from pyroxene granulites
Location of the samples given in Appendix

Analyst: M.Q.J.

Anal. SH 16 taken from
Jan and Kempe, 1973
(Anal. C.J. Elliot, V.K. Din, and A.J. Easton).

† $mg = 100Mg/(Mg+Fe^{2+}+Fe^{3+}+Mn)$

and Smith (1966) but the Ti and Ca contents of the Kohistan pyroxenes are distinctly higher. Although similar values for the two elements have been reported for some granulite facies orthopyroxenes (Binns, 1965; Davidson, 1968), the presence of clinopyroxene impurities, mainly as exsolution lamellae, is a contributory factor here.

The values for Cr, Ni, and Pb are much higher and those of Zn and Li lower in the orthopyroxene from hornblende websterite granulite (US 24). This rock is not a part of the pyroxene granulites but forms an independent mass in the amphibolites to the south. Of the two hypersthenes from dioritic rocks, no. 647 bears a close resemblance to those from the granulites except for its slightly lower Al, Cr, Ni, and V. Jan (1977*b*) has suggested that such hypersthene diorites may be retrograde granulites and it is possible that the hypersthene relics have been depleted in certain elements, perhaps by ionic diffusion. The neighbouring hornblende and biotite in 647, probably formed mainly at the expense of hypersthene and plagioclase, are considerably richer in Cr, Ni, and V than the hypersthene. The orthopyroxene in quartz diorite 448 is distinctly different from those of the granulites. It is richer in Ti, Na, K, Li and, especially, Ca and Cr, and poorer in Co. The higher CaO may be due to its formation at a higher temperature than the granulite orthopyroxenes, and imperfect separation of finely exsolved lamellae of clinopyroxene. However, petrographic evidence shows that the host rock 448 is a much younger and shallower intrusion in metasediments to the north of the granulite belt, and is not comagmatic with the granulites.

Compared with the bronzite and hypersthene analyses from the Bushveld complex (Atkins, 1969, Table I, anal. 6a and 8a), the Swat hypersthenes have considerable Zn (not detected in those from Bushveld) and Cu, slightly higher Co and Cr, similar V, but distinctly lower Ni. Compared with the orthopyroxenes from the three intermediate rocks and one basic rock of the Madras charnockite series (Howie, 1955, Table IX) the values for Li, Pb, Co, and V cover similar ranges, whilst the Swat orthopyroxenes are richer in Ni.

The CaO content of orthopyroxene has long been considered to increase with temperature (Atlas, 1952; Kuno, 1954; Hess, 1960). Binns (1962) found that orthopyroxenes in the higher temperature zones of the Broken Hill area of New South Wales have higher CaO than in the low-temperature zone, although Binns (1965) also considered that CaO variation in the orthopyroxenes may be more susceptible to the influence of other components. Davidson (1968) found a lower miscibility gap in the coexisting pyroxenes of the Quairading area when compared to those of Broken Hill; the former being thought to have formed at a higher metamorphic grade. Recent experimental data support the idea that the amount of diopside in enstatite, and vice versa, increases with temperature. The influence of pressure is variously interpreted, but is generally considered to have an opposite effect to that of temperature (Lindsley and Dixon, 1976; Akella, 1976; Nehru, 1976).

A comparison of igneous and metamorphic orthopyroxenes (cf. Deer *et al.* 1978, Tables 3 and 4) shows that the latter, in general, are more aluminous. The Al_2O_3 content (1.3–3.1 %) of the Swat granulite orthopyroxenes is well within the range of those from granulites and none of them is exceptionally high in Al_2O_3. However, their alumina contents, in general, are higher than those of similar hypersthenes (En_{60-50}) and Ca-poor clinopyroxenes of the Skaergaard, Stillwater, and Bushveld complexes. The amount of Al_2O_3 in orthopyroxene depends upon the bulk composition of the rock (Howie, 1965; Binns, 1965; Davidson and Mathison, 1974), pressure and/or temperature of formation (Hess, 1952; Eskola, 1957; Green, 1964; Dobretsov *et al.* 1972; Medaris, 1972), and the coexisting minerals (Binns, 1965; Windley and Smith, 1974).

Experimental studies in the $MgO-Al_2O_3-SiO_2$ system confirm that Al_2O_3 in enstatite increases with temperature. The dependence of the amount of alumina in orthopyroxene on pressure, however, is complex; the problem has been discussed in some detail by Dobretsov *et al.* (1972). Boyd and England (1960) found that increasing pressure at a given temperature increases the solubility of Al_2O_3 in enstatite. However, in spinel- and garnet peridotites (MacGregor, 1974), and in garnet websterite compositions (Akella, 1976), a fall in the Al content of the enstatite with increasing pressure has been noted. Obata (1976) found that Al_2O_3 in orthopyroxene is much less sensitive to pressure than suggested by MacGregor (1974) and in spinel peridotite at temperatures below 1000 °C is essentially a function of temperature. However, in plagioclase peridotite, according to Obata, the amount of alumina in orthopyroxene is very sensitive to pressure, and in spinel pyroxenite to both pressure and temperature. Presnall (1976) reported that the Al_2O_3 isopleths of enstatite in the spinel lherzolite field have gentle and probably negative dT/dP slopes in sharp contrast to the steep and positive slopes reported by MacGregor (1974). The addition of other components further complicates the situation. Boyd (1970) and Green and Ringwood (1970) noted that in natural multicomponent systems containing also Ca and Fe^{2+}, the orthopyroxenes in equilibrium with garnet are considerably poorer in alumina than in the pure $MgO-Al_2O_3-SiO_2$

system. Most of the above data concern simple systems usually involving garnet and/or spinel and cannot, therefore, be meaningfully applied to the Swat hypersthenes. It can, however, be said that P–T estimates based solely on the alumina content of the orthopyroxene can be misleading.

Clinopyroxenes. Chemical analyses of twenty-one clinopyroxenes, nineteen of them coexisting with the orthopyroxenes discussed above, are presented in Table II. Analysis 324 is of a light-green sahlite from a calcareous horizon and US 24 from the hornblende websterite granulite, both occurring within rocks from the southern amphibolite belt. 448 is from a quartz diorite to the north whilst the remaining clinopyroxenes are from the pyroxene granulites.

The granulite clinopyroxenes have dark-green powders and are green and weakly pleochroic in thin section, the green colour approximately matching the γ absorption of the coexisting orthopyroxene. Their density and refractive indices increase with increase in iron content. Exsolved blebs of orthopyroxene parallel to (100), and blades of a black to reddish-brown material are found in many cases, as in the coexisting orthopyroxenes. Some workers have described brown blades of Ti-bearing minerals in pyroxenes but these are not found here. Analysis US 9 is of a clinopyroxene free of the brownish material and US 9x of the same clinopyroxene contaminated with the brownish material. It can be seen that the essential difference in the chemistry of the two fractions is in FeO rather than in Ti or Mn. It is therefore probable that the brownish blades in the clino- and orthopyroxenes of the Kohistan granulites are hematitic in nature and exsolved during metamorphism, possibly under increased f_{O_2}.

In all the analyses except 324, the amount of Al^{iv} exceeds Al^{vi}. With increase in Fe, the clinopyroxenes generally become poorer in Si, Ca, Cr, and Ni. The compositions of the two sahlites (one optically determined) from the 'amphibolites' are more calcic than those from the granulites (fig. 1). The plot of the latter occupies a limited area in the augite–sahlite fields, the range of Mg being 35–39.4, Fe 15.6–20.0, and that of Ca 46.5–43.2. An optically determined composition from a pyroxenite layer in granulite (SK 354) is $Mg_{35}Fe_{17}Ca_{48}$. The analysis of clinopyroxene 221 ($Mg_{37.2}Fe_{23.7}Ca_{39.1}$) plots away from the rest; its host rock comes from the northernmost part of the area and its relationship with the remaining granulites is not clear. Although the close clustering of the analyses is probably largely the result of similar bulk chemistry and the amount of associated plagioclase and hypersthene, uniformity of metamorphic conditions may also have played some role in producing compositionally similar clinopyroxenes. For example, rock 322 has a much higher Niggli mg and a higher Mg/Ca ratio than the rest of the granulites from which clinopyroxenes have been analysed (Jan, 1977*b*), yet clinopyroxene 322 has a similar composition to the others.

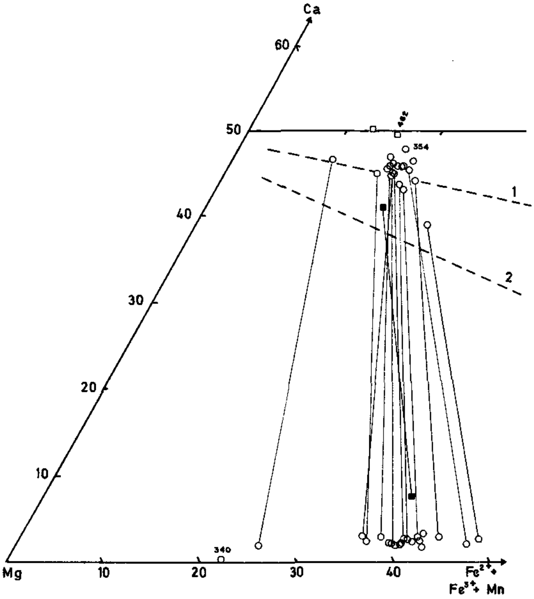

FIG. 1. A portion of the pyroxene trapezohedron showing the Mg-($Fe^{2+}+Fe^{3+}+Mn$)-Ca plots for the forty-four pyroxenes of Kohistan from the granulites (open circles), quartz diorite 448 (filled squares), and from the amphibolite belt (open squares). Orthopyroxene 647 has been included in the granulite pyroxenes. Numbers indicate optically determined compositions. Tie-lines have been drawn between some pairs only for the sake of clarity. Broken lines 1 and 2 show, respectively, the trends of granulite and igneous clinopyroxenes coexisting with Ca-poor pyroxene (after Binns, 1965).

By plotting seventy-seven clinopyroxenes coexisting with orthopyroxenes, Binns (1965) showed that metamorphic clinopyroxenes contain more Ca than those crystallized from slowly cooled magmas; this was interpreted as being due to low temperature of formation of the metamorphic rocks. The Swat granulite clinopyroxenes also plot along the metamorphic rather than igneous trend considered by Binns. Compared with the augitic pyroxenes from the Skaergaard (Brown, 1957), Stillwater (Hess, 1960), Bushveld (Atkins, 1969; Buchanan, 1979) and Dufek, Antarctica (Himmelberg and Ford, 1976) intrusions, considered to have differentiated from tholeiitic magmas, the Swat granulite clinopyroxenes are richer in Ca and, in general, also in Al and Na. It would seem as if some

TABLE II Analyses of the clinopyroxenes

	US 24	SI 324	SK 448	SI 192	SK 600	SK 615	SK 422	SK 528	SH 16	SK 592	US 15	SI 221	SK 322	US 9	US 9X	SI 197	SI 188a	SI 188b	SI 193	US 14	SI 182
SiO_2	52.09	53.34	51.47	51.23	50.73	51.58	51.12	50.38	51.16	50.56	50.57	51.10	50.14	50.68	-	51.11	50.36	50.56	51.01	50.97	49.61
TiO_2	0.56	0.15	0.63	0.40	0.45	0.37	0.34	0.46	0.43	0.55	0.32	0.66	0.61	0.56	0.49	0.41	0.56	0.40	0.29	0.37	0.51
Al_2O_3	3.53	0.71	2.28	2.69	3.08	2.26	2.80	3.89	4.09	3.34	2.91	2.25	4.37	3.49	3.40	2.68	3.70	3.69	2.29	2.59	4.50
Fe_2O_3	1.48	0.49	1.29	2.14	2.63	2.11	2.58	3.04	2.16	1.90	2.63	2.51	3.02	2.56	2.61	2.12	2.77	2.52	2.34	2.82	2.79
FeO	4.71	6.56	9.48	7.42	7.30	8.27	7.77	8.02	7.54	8.50	8.97	11.64	6.94	7.76	8.81	8.33	7.81	8.04	8.82	9.23	8.24
MnO	0.18	0.78	0.33	0.24	0.25	0.31	0.33	0.28	0.22	0.27	0.31	0.36	0.27	0.26	0.28	0.28	0.31	0.35	0.37	0.33	0.32
MgO	14.85	12.97	13.79	13.60	12.97	13.02	12.96	12.60	12.42	12.66	12.70	12.55	12.40	12.54	12.83	12.62	12.19	12.12	12.18	12.17	11.54
CaO	22.32	24.36	19.38	21.63	21.62	21.37	21.47	20.32	20.98	21.09	20.40	18.35	21.46	21.60	20.78	21.90	21.41	21.37	21.53	20.93	21.42
Na_2O	0.51	0.35	0.37	0.48	0.52	0.49	0.52	0.66	0.61	0.50	0.64	0.42	0.68	0.59	0.57	0.50	0.55	0.58	0.51	0.68	0.67
K_2O	0.00	0.03	0.05	0.01	0.00	0.01	0.00	0.01	0.02	0.01	0.02	0.02	0.01	0.00	0.01	0.00	0.00	0.01	0.01	0.01	0.01
P_2O_5	0.00	0.03	0.07	0.00	0.00	0.03	0.04	0.00	-	0.00	0.06	0.00	0.00	0.00	0.00	0.00	0.01	0.03	0.30	0.01	0.00
TOTAL	100.23	99.77	'99.46'	99.84	99.55	99.82	99.93	99.66	'99.96'	99.55	'99.69'	99.86	99.90	100.04	-	99.95	99.67	99.67	99.65	100.11	99.61
α	1.683	-	-	1.687	-	-	1.688	-	-	-	1.691	-	-	1.689	-	-	-	-	-	1.691	-
β	1.690	-	-	1.695	-	-	1.697	-	-	-	1.698	-	-	1.698	-	-	-	-	-	1.699	-
γ	1.709	-	-	1.713	-	-	1.714	-	-	-	1.716	-	-	1.717	-	-	-	-	-	1.719	-
γ:z	42°	-	-	44°	-	-	43°	-	-	-	44°	-	-	44°	-	-	-	-	-	44°	-
D	3.336	3.360	3.368	3.372	3.375	3.377	3.372	3.390	-	3.366	3.402	-	3.363	3.380	-	3.384	3.382	3.369	3.386	3.405	3.400
NUMBERS OF IONS ON THE BASIS OF SIX OXYGENS																					
Si	1.911	1.992	1.939	1.914	1.904	1.934	1.914	1.890	1.898	1.904	1.909	1.930	1.874	1.896		1.917	1.892	1.900	1.930	1.917	1.871
Al	0.089	0.008	0.061	0.086	0.096	0.066	0.086	0.110	0.102	0.096	0.091	0.070	0.126	0.104		0.083	0.108	0.100	0.070	0.083	0.129
Al	0.064	0.023	0.040	0.033	0.040	0.034	0.038	0.062	0.077	0.052	0.038	0.030	0.067	0.050		0.036	0.056	0.063	0.032	0.032	0.071
Ti	0.015	0.004	0.018	0.011	0.013	0.010	0.010	0.013	0.012	0.016	0.009	0.019	0.017	0.016		0.012	0.016	0.011	0.008	0.011	0.015
Fe^{3+}	0.041	0.014	0.037	0.060	0.074	0.060	0.073	0.086	0.060	0.054	0.074	0.071	0.085	0.072		0.060	0.078	0.071	0.067	0.080	0.079
Fe^{2+}	0.145	0.205	0.299	0.232	0.229	0.259	0.243	0.252	0.234	0.268	0.283	0.368	0.217	0.243		0.261	0.245	0.253	0.279	0.290	0.260
Mn	0.006	0.025	0.011	0.008	0.008	0.010	0.011	0.009	0.007	0.009	0.010	0.012	0.009	0.008		0.009	0.010	0.011	0.012	0.011	0.010
Mg	0.812	0.722	0.774	0.758	0.726	0.728	0.723	0.705	0.687	0.711	0.715	0.707	0.691	0.699		0.706	0.683	0.679	0.687	0.682	0.649
Ca	0.877	0.975	0.782	0.866	0.869	0.858	0.861	0.817	0.834	0.851	0.825	0.743	0.860	0.866		0.880	0.862	0.861	0.873	0.844	0.866
Na	0.036	0.025	.027	0.035	0.038	0.036	0.038	0.048	0.044	0.036	0.047	0.031	0.049	0.043		0.036	0.040	0.042	0.037	0.050	0.049
K	0.000	0.001	0.002	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.001	0.001	0.000	0.000		0.000	0.000	0.000	0.000	0.000	0.000
X + Y	1.996	1.994	1.990	2.003	1.997	1.995	1.997	1.992	1.96	1.997	2.002	1.982	1.995	1.997		2.000	1.990	1.991	1.995	2.000	1.999
Mg	43.20	37.2	40.7	39.4	38.1	38.0	37.8	37.7	37.7	37.6	37.5	37.2	37.1	37.0		36.8	36.3	36.2	35.8	35.8	34.8
Fe	10.15	12.6	18.2	15.6	16.3	17.2	17.1	18.6	16.5	17.4	19.3	23.7	16.7	17.1		17.3	17.8	17.9	18.7	20.00	18.7
Ca	46.65	50.2	41.1	45.0	45.6	44.8	45.1	43.7	45.8	45.0	43.2	39.1	46.2	45.9		45.9	45.9	45.9	45.5	44.2	46.5
PPM																					
Co	36	107	47	50	50	109	50	52	-	48	52	56	27	48		52	54	50	45	52	46
Cr	2110	17	375	204	169	24	34	132	1000	150	84	109	83	248		127	59	15	32	105	133
Cu	7	78	96	25	20	21	52	51	-	20	65	124	45	22		38	280	133	105	64	14
Li	5	7	24	10	10	6	11	12	-	8	15	11	5	10		13	10	10	10	17	9
Ni	229	441	106	148	117	100	86	121	100	87	97	64	58	140		106	63	51	61	116	73
Pb	<5	58	20	<5	<5	<5	11	10	-	<5	<5	<5	12	5		<5	<5	<5	<5	<5	<5
V	232	6	322	308	398	105	294	411	800	427	317	406	245	439		393	379	304	317	355	372
Zn	57	700	-	122	101	443	163	124	100	92	153	155	-	134		127	253	168	153	178	137

Analyst: M.Q.J.

SH 16 taken from Jan and Kempe, 1973
(Anal. Elliot and others (total includes 0.33 % H_2O)

* US 24: From websteritic granulite in amphibolites
SI 324: From a calc-silicate horizon in amphibolites
SK 448: From a quartz diorite (total includes 0.32 % H_2O)
The remaining analyses are from the pyroxene granulites

US 9X is similar to US 9 but for contamination by a brown iron oxide blades
Location of the samples given in Appendix
1. includes 0.11 % H_2O

plagioclase component has been 'accommodated' in clinopyroxene solid solution during metamorphism.

Experimental data referred to in the section on orthopyroxenes suggest that with increasing temperature the CaO content of the clinopyroxene rapidly falls due to a greater solid solution of orthopyroxene. On the other hand, exsolution of ortho- in clinopyroxene falls with increasing pressure although the influence of pressure is negligible under 1000 °C (Warner and Luth, 1974; MacGregor and Basu, 1976; Mori and Green, 1976). It also appears that jadeite and, within certain limits, tschermaks component, increase with pressure. Thus it would seem probable that the higher Ca, Na, and Al in the Swat granulite clinopyroxenes are suggestive of higher pressure and lower temperature conditions than those operating during the crystallization of deep-seated tholeiitic magmas.

The trace elements of the granulite clinopyroxenes have been compared with the augites from the Bushveld complex (Atkins, 1969) and the norite and intermediate members of the Madras charnockite series (Howie, 1955). The Bushveld augites of similar Mg:Fe ratio as those of the Swat clinopyroxenes are poorer in Cr, Ni, Cu and, especially, V and Zn (the latter was not detected in the Bushveld augites), while Co values are similar. The Madras pyroxenes are richer in Li and a little poorer in Ni; Co and Cr have similar ranges, but V has a much greater range (50–400 ppm) compared to those of Swat (245–440 ppm, ignoring the lowest and the highest values).

The clinopyroxene from quartz diorite (448), like the coexisting orthopyroxene, is significantly different from those of the granulites. It is much lower in Ca, which may be due to its formation at higher temperature although contamination of finely exsolved orthopyroxene is also a contributory factor. It is also lower in Fe_2O_3, Na_2O, and higher in TiO_2 and, especially, in Cr, Li, and Pb. The two clinopyroxenes from the amphibolite facies rocks are characteristically richer in Ca. The composition $Mg_{35}Fe_{15.5}Ca_{49.5}$ of one of these (from amphibolite 462) has been determined from β and 2V. The other (from a calc-silicate rock 324), analysed chemically, is also a sahlite ($Mg_{37.2}Fe_{12.6}Wo_{50.2}$). The significant features of this clinopyroxene are the lower Ti, Fe^{3+}, Cr and, particularly, V contents, with higher Mn, Co and, especially, Ni, Pb, and Zn. The clinopyroxene from the websterite 24 in amphibolites has higher Ni, and Cr, and lower Cu, Zn, and V than those from the granulites.

Pyroxene tie-lines. Hess (1941) pointed out that the tie-lines of the coexisting igneous pyroxenes intersect the Wo–En join at around Wo_{75}. Based on optical data, Wilson (1960) found different intersections and thought that intersections towards the Mg end reflect lower grade metamorphic rocks either remobilized or pyroxene granulites close to the boundary with hornblende granulites. Later work by Subramaniam (1962) did not contradict this view. The problem of tie-lines has also been studied by Muir and Tilley (1958), O'Hara (1960), and Brown (1961); the last author pointed out the discrepancies in composition determined from optical measurements and chemical analyses.

Brown (1961) showed that the average point of intersection on the Wo–En join for igneous and metamorphic pyroxenes is the same, at Wo_{82}, and suggested that the exact position of this point has no real significance in crystal chemistry or petrogenesis: it is merely indicative of equilibrium conditions (Brown, 1962). Values departing greatly from Wo_{60} to Wo_{90}, in the writers' opinion, indicate that either the pyroxenes are not in equilibrium, or that the composition of one or both the pyroxenes has been wrongly determined.

In the case of the Swat granulite pyroxenes (including US 24), fourteen of the tie-lines intersect the Wo–En join between $Wo_{80.5}$ and $Wo_{76.5}$ and the remaining four between Wo_{85} and Wo_{75}. Useful conclusions cannot be drawn from these intersections but the close clustering of the points may be suggestive of the attainment of uniform equilibrium conditions.

Mg–Fe–Mn distribution. The Fe^{2+}–Mg distribution in minerals, especially in pyroxene pairs, has been investigated in detail by a number of workers (Mueller, 1960, 1961; Kretz, 1961, 1963; Bartholomé, 1961, 1962). The distribution coefficient (K_D) for the exchange reaction is defined as:

$$K^{A-B}_{D_{\text{Mg-Fe}}} = \frac{X^{A}_{\text{Mg}}(1-X^{B}_{\text{Mg}})}{X^{B}_{\text{Mg}}(1-X^{A}_{\text{Mg}})}$$

where X_{Mg} refers to mole fraction

$$\frac{\text{Mg}}{\text{Mg}+\text{Fe}^{2+}} \text{ in minerals A and B.}$$

Thus $K^{A-B}_{D_{\text{Mg-Fe}}} = (\text{Mg}/\text{Fe}^{2+})^{A}/(\text{Mg}/\text{Fe}^{2+})^{B}$.

Kretz thought that K_D changes with pressure and temperature but is unaffected by variations in Fe:Mg ratio under a particular set of equilibrium conditions. Mueller (1961) considered that the presence of other components will either not appreciably affect K_D or will affect equally both nominator and denominator. Kretz (1963) studied a number of igneous and metamorphic pyroxene pairs and found that the average value of K_D for the former is distinctly higher than that for the latter. On theoretical grounds, K_D should increase with temperature or ln K_D should approach zero. (The K_D value has been expressed in an inverted form by Bartholomé (1961, 1962) and Binns (1962) and this has been followed by a number of other workers. In this form, a K_D value of 0.54 (Madras charnockites) becomes 1.85 and the typical igneous value of 0.73 becomes 1.37).

Ghose (1965) studied the structure of a granulite facies orthopyroxene ($Fs_{53.5}$) and found it to be completely ordered, with Mg ions preferring the $M1$ site and Fe^{2+} ions occupying the $M2$ site. Further investigations of the orthopyroxene structure have confirmed the suggestion that Mg ions show a strong preference for the $M1$ site and Fe^{2+} ions for the $M2$ site (cf. Kretz, 1972). Lindh (1974) found that Fe^{2+} prefers the $M2$ position in clinopyroxene and has a temperature dependent preference for $M2$ positions in orthopyroxene. The $M2$ preference may be ascribed to either or both of two factors, ionic radius and crystal-field stabilization energy.

Binns (1962) showed that $K_D^{\text{Mg-Fe}}$ is not independent of composition and decreases in iron-rich pyroxene pairs. More recently, Davidson (1968, 1969) showed that $K_D^{\text{Fe-Mg}}$ in pyroxene pairs from the Quairading granulites, Western Australia, remains constant for more magnesian pyroxenes but falls rather rapidly in those pairs whose orthopyroxene is less magnesian than Fe_{40}. Since the rocks of these pyroxene pairs seem to have formed within a narrow temperature range, Davidson concluded that the pyroxenes may express a non-ideal distribution of Fe and Mg. A similar decrease in K_D^{Fe} at Fs less than 41–43 was noted by Scharbert and Kuart (1974) for garnet-orthopyroxene pairs from lower Austrian granulites (their calculated K_D involves Ca in addition to Fe^{2+} and Mg). Virgo and Hafner (1970) showed that Fe^{2+} and Mg are ideally distributed at each site between En_{100} to En_{40}; that less magnesian pyroxenes are non-ideal mixtures; and that non-ideal mixing at the sites decreases with temperature.

Ray and Sen (1970) found that Ca content of the clinopyroxene exerts a considerable influence on the K_D. The statistical study of pyroxenes by Lindh (1975) has shown that K_D is influenced by the Fe/Mg ratio, by the Ca content, especially of clinopyroxene, and by the content of Al^{iv}. In the Precambrian Highlands of New Jersey, the $K_D^{\text{Fe-Mg}}$ increase with increasing pyroxene wt% FeO, the opposite relationship to that found by Davidson (1968). Thus, Maxey and Vogel (1974) concluded that K_D^{Fe} 'is more apt to increase with decreasing pyroxene FeO* (increasing MgO) in suites of relatively Fe-rich pyroxenes, whereas an opposite type trend is more likely to be present in suites of relatively Fe-poor pyroxenes'. They also noted that since the Fe–Mg content of the pyroxenes shows sympathetic relationship with the FeO/MgO ratio of the host rocks, the observed change in K_D in New Jersey amphibolites, consequently, reflects a variation in rock composition and not in crystallization temperature.

The problem has also been investigated by Fleet (1974*a*, *b*), using a different approach where it is not necessary to assume any exchange reaction. He considered that the partition of Mg, Fe^{3+}, and Mn correlates with inferred cooling rates, the partition coefficients of pyroxenes from extruded and other quickly cooled rocks most nearly approaching unity. According to Fleet (1974*c*), 'occupancy of Mg and Fe^{2+} in the $M2$ site of the Ca-rich pyroxene effectively compensates for the expected variation in K_D with composition resulting from intracrystalline partition in Ca-poor pyroxenes, and this largely accounts for the difference in K_D between igneous and metamorphic pyroxenes'.

But despite the various chemical factors that can influence K_D, it is commonly agreed that crystallization conditions (mainly temperature) have significant control over K_D (Kretz, 1963; Scharbert and Kuart, 1974; Sen and Manna, 1976). Although there is a scatter in K_D values, statistically igneous pyroxenes have considerably different higher $K_D^{\text{Mg-Fe}}$ values than those of metamorphic pyroxenes (Lindh, 1975).

The $K_D^{\text{Mg-}Fe^{2+}}$ and $K_D^{\text{Mn-Mg, }Fe^{2+}}$ for the Swat pyroxene pairs are presented in Table III. The average $K_D^{\text{Mg-Fe}}$ for the pyroxene granulites (0.57) is very close to that (0.54) for metamorphic pyroxenes (Kretz, 1963). The narrow range in K_D values (0.52–0.61) may suggest a narrow range of temperature in addition to similarity of composition. The K_D (0.62) for the quartz diorite 448 is significantly higher than the average value for the pyroxene granulites. The K_D values for the granulite pyroxenes were plotted against various cations in ortho- and clinopyroxenes to make a visual assessment of compositional control over them. The $K_D^{\text{Fe-Mg}}$ shows a general decrease with increase in Al and Al^{iv} in clinopyroxene and, interestingly, with Ti in ortho- as well as clinopyroxene. The K_D plots against the rest of the cations and against Fe/Mg in ortho- and clinopyroxene, however, are scattered and inconclusive.

Although some workers (e.g. Scharbert and Kuart, 1974; Sen and Manna, 1976) think that K_D^{Mn} is highly sensitive to temperature and decreases with increase in metamorphic grade, Kretz (1963) and Lindh (1974) did not find appreciable difference in the K_D^{Mn} values for igneous and metamorphic pyroxenes. The K_D^{Mn} in the Swat pyroxenes also shows no appreciable dependence on composition, at least as far as the Mg/Fe ratio in the pyroxenes is concerned. The average value (1.13) for the five Mg-rich pyroxene pairs from the ultramafic rocks of Jijal (Jan and Howie, in preparation) is very close to that (1.10) of the hypersthene–augite/sahlite pairs from the pyroxene granulites.

The pyroxenes show a limited iron enrichment (fig. 1) and in this and other respects are comparable with those of the younger volcanic islands of Tonga, southwestern Pacific (Ewart *et al.*, 1973); the coexisting pyroxenes exhibit equilibrium distributions of Mg, Mn, V, and Ni. Comparison may also be made with the clinopyroxenes of the ophiolitic basalts with 'oceanic' characters from the western Mediterranean (Capedri and Venturelli, 1979). The major element compositional ranges are similar and the Swat clinopyroxenes also show a high Ti–Al correlation; their Ti/Al ratio of around 0.1 is comparable with that of clinopyroxenes from

TABLE III. *Distribution coefficients of the Swat pyroxene pairs*

	X_{opx} = Mg/(Mg+Fe^{2+})	X_{cpx} = Mg/(Mg+Fe^{2+})	$K_D^{Mg\text{-}Fe}$	X_{opx} = Mn/(Mn+Mg+Fe^{2+})	X_{cpx} = Mn/(Mn+Mg+Fe^{2+})	$K_D^{Mn\text{-}Mg,Fe}$
SH 16	0.6400	0.7459	0.606	0.0086	0.0075	1.148
US 24	0.7277	0.8485	0.477	0.0058	0.0062	0.935
US 9	0.6229	0.7420	0.574	0.0091	0.0084	1.084
US 14	0.5854	0.7016	0.600	0.0120	0.0112	1.072
US 15	0.5876	0.7164	0.564	0.0107	0.0099	1.082
SI 182	0.5910	0.7140	0.579	0.0117	0.0109	1.074
SI 188a	0.6080	0.7360	0.556	0.0112	0.0107	1.047
SI 188b	0.5997	0.7285	0.558	0.0131	0.0117	1.121
SI 192	0.6621	0.7657	0.600	0.0087	0.0080	1.088
SI 193	0.5616	0.7112	0.520	0.0144	0.0123	1.173
SI 197	0.6043	0.7301	0.565	0.0109	0.0092	1.187
SI 221	0.5316	0.6577	0.591	0.0145	0.0136	1.067
SK 322	0.6594	0.7610	0.608	0.0090	0.0092	0.978
SK 422	0.6188	0.7484	0.546	0.0122	0.0113	1.081
SK 448	0.6151	0.7213	0.618	0.0112	0.0097	1.156
SK 528	0.6279	0.7367	0.603	0.0095	0.0092	1.033
SK 592	0.6114	0.7263	0.593	0.0102	0.0087	1.174
SK 600	0.6334	0.7602	0.545	0.0093	0.0082	1.135
SK 615	0.6089	0.7376	0.554	0.0118	0.0098	1.206
Average (excluding quartz diorite 448)			0.569	—	—	1.094

the Pindos Massif, Greece. In terms of the Na_2O–TiO_2–MnO relationships used by Nisbet and Pearce (1977) to define clinopyroxene compositions from different tectonic settings, the Swat clinopyroxenes define a trend typical of those from island arc basalts.

P–T conditions of the Swat pyroxene granulites. Because of their unzoned nature and apparent attainment of equilibrium, the Swat pyroxenes are suitable for estimation of metamorphic conditions. Temperature estimates according to eight different methods are presented in Table IV. Although Hewins (1975) and Akella (1976) found the Wood–Banno (1973) method satisfactory for rocks that have not formed at very high pressure, a number of workers (Fleet, 1974*c*; Mori, 1977; O'Hara and Yarwood, 1978) feel that the method needs revision. Wood (1975) himself suggested a downward revision of 60 °C for temperatures obtained by this method, and it appears that for the Swat pyroxene compositions the method of Wells (1977) also produces temperature overestimates of 80°–100 °C (see also Weaver *et al.*, 1978). On Fleet's (1974*c*) diagram the Swat clinopyroxenes coexisting with orthopyroxenes plot between the 900 °C isotherm and the clinopyroxenes from charnockites and granulites, for which he estimated a temperature of 750 °C. It is concluded that the Swat granulites crystallized at temperatures of around 800 °C.

Two significant points emerge from the geothermometry: (1) the temperature estimates are within the limits of the granulite facies conditions but considerably lower than those of the magmatic pyroxenes, and (2) the estimates by any one method are within a range of 110 °C, ignoring specimen 221 whose dubious status has already been pointed out. The narrow temperature range is in accordance with the narrow ranges in the composition of the pyroxenes, their tie-line intersections, and their $K_D^{Mg\text{-}Fe}$ values. Considering the vastness of the granulite terrain, such a uniformity of metamorphic conditions is a little surprising.

Assuming an 800 °C temperature, a pressure of 7–8 kbar is estimated from the two-feldspar geothermometer of Stormer (1975). Experimental work by Green and Ringwood (1967) suggests that pressures in excess of 8 kbar are needed for the appearance of garnet at 800 °C. Similarly, according to the data of Hensen (1976), pressures of 9–10 kbar are needed at 800 °C for the appearance of garnet from the reaction between orthopyroxene and plagioclase (An_{50}) in composition with Mg/(Mg+Fe) = 0.50 (close to the Swat granulites; Jan, 1977*b*). Garnet, along with amphibole, is present in only one place (rock S I 217, 3.5 km north of Seo, Indus Valley) and although its formation may in part be due to the olivine-normative nature of its host rock, its presence here and in a few pegmatitic rocks in that area cannot be completely divorced from physical environments. The absence of olivine even

TABLE IV. *Temperature estimates (°C) for the pyroxene granulites*

	Wood and Banno, 1973	Perchuk and Vaganov, 1977	Wells, 1977	Others	
US 9	952°	770°	869°		
14	854°	695°	892°	670° at 7.5 kbar	(Stormer, 1975) Two-feldspar geothermometer
15	884°	695°	935°	800° at 7.5 kbar	
SH 16	894°	750°	914°		
SI 182	819°	675°	839°		
188a	852°	710°	860°		
188b	850°	700°	876°		
192	886°	795°	905°	700°,* 780°,† 800°,† 700°,†† 750°†† 740°,†† 730°††	
193	829°	665°	859°		
197	839°	705°	860°		
221	?937°	?600°	1022°		
SK 322	862°	775°	870°		
422	865°	730°	892°		
528	907°	755°	939°		
592	876°	735°	911°		
600	866°	750°	885°		
615	872°	725°	904°		
SI 217				770°*	
Average without 221	863°	727°	889°		

Approximate average temperature by the method of Fleet (1974*c*) = 800 °C.

Wood and Banno (1973), Perchuk and Vaganov (1977), and Wells (1977) methods based on compositions of coexisting pyroxenes.

* $K_{D_{\text{Ca-Na,K}}}^{\text{Amph-Plg}}$ (Perchuk, 1966).

† Mg–Fe distribution in amphibole and clinopyroxene/orthopyroxene (Perchuk, 1969).

†† Mg–Fe distribution in biotite and orthopyroxene/clinopyroxene (Perchuk, 1969).

Although the latter values are in accordance with others, the biotite cross-cuts other minerals and looks secondary. Either the four estimated temperatures are much too fortuitous or, despite an opposing textural evidence, the biotite has reached a physicochemical equilibrium with the coexisting pyroxenes.

in the olivine-normative rocks may suggest pressures in excess of 6 kbar at 800 °C (Dobretsov *et al.*, 1972). Dobretsov *et al.* (1973) consider that the two-pyroxene gneiss (granulite) facies rocks crystallize generally between 750° and 900 °C and 6 and about 12 kbar. It is thus reasonable to assume that pressures of 7 to 8 kbar were operating during the recrystallization of the pyroxene granulites. Estimated temperatures for US 24 (metamorphosed hornblende websterite in amphibolite) are 860 °C (amphibole-clinopyroxene), 845 °C (two-pyroxenes), and 890 °C (orthopyroxene-amphibole) by the methods of Perchuk (1969), and 934 °C by the model of Wood and Banno (1973). All these estimates are much higher than those for the neighbouring amphibolites and cast doubt on the *in situ* metamorphism of the rock; it may have been emplaced tectonically or remobilized from depth after the main amphibolite facies metamorphism.

Temperature estimates for the quartz diorite 448 are 920 °C (Fleet, 1974*c*), 905 °C (Wells, 1977), and 976 °C (Wood–Banno). Three estimates based on biotite-pyroxene and the two-pyroxene methods of Perchuk are significantly lower (725°–800 °C) and are discarded. The lower values obtained by biotite-pyroxene methods may also not be reliable because the pyroxenes do not seem to be in equilibrium with the biotite. Temperatures of more than 900 °C are in accord with the small size and subporphyritic intrusive nature of the quartz diorite.

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Appendix 1

Analytical techniques. Minerals were separated by magnetic separator and heavy liquids from rocks in which the two pyroxenes showed equilibrium texture and, as with the plagioclase, little or no zoning. Small volumes of rocks, weighing 500 g or less, were taken for separation. The analysed concentrates appear to be > 99 % pure, except those from 448 (~ 97 % pure), the impurities being the other pyroxene, usually as lamellae, and blades, and rare granules of iron oxide.

The pyroxenes were analysed by combining classical and instrumental methods. SiO_2 was determined gravimetrically; FeO by titration (Wilson method); TiO_2 and P_2O_5 colorimetrically, following Riley and using a Unicam SP 600 series 2 spectrophotometer; and H_2O (in only a few cases) by the modified Penfield method. The rest of the elements were measured by atomic absorption Unicam SP 1900, using the emission mode for the alkalis. In the case of total Fe_2O_3, MgO, and CaO, 1 % $LaCl_3$ solution was added to the diluted stock solution, and for Al_2O_3 an equivalent amount of Na solution was added. Standards of different laboratories were used for reference. The reported values are the averages of two or more runs on the same stock solution and should be accurate to ± 1 % of the amounts present for the major elements. Compositions of three more pyroxenes were determined from β and 2V. Refractive indices were determined by the dispersion method at room temperature; and density by pycnometer, recalculated to 0°C.

Appendix 2

Location and mode of the host rocks

In the following is given the location and other relevant information for the rocks from which pyroxene compositions have been determined; the samples are arranged in numerical order. The distances, from towns/landmarks up to the sample site, were directly measured with a scale after the samples were plotted on a map. Abbreviations for mineral names are: Am = amphibole, Ap = apatite, Bio = biotite, Cpx = clinopyroxene, Epi = epidote, Hbl = hornblende, Kf = K-feldspar, Ore = opaque minerals, Opx = orthopyroxene, Plg = plagioclase, and Qz = quartz.

US 9. Pyroxene granulite from a lens in amphibolites, 2.5 km SW of Fatehpur and to the S of Jabrai, along road W of river, Upper Swat. Contains 75.2 % Plg (An$_{50}$) 14.2 % Opx, 6.0 % Cpx, 2.5 % Qz, 1.2 % Ore, and 0.9 % Ap, Am, Bio.

US 14. Pyroxene granulite just N of Bartana, about 16 km NW of Matta, Upper Swat. Contains Plg, Opx, Qz, Cpx, Kf, Ore, Hbl, Ap, and Bio. Composition of Plg (An$_{46}$) and Kf (Or$_{86}$ Plg$_{14}$) determined from refractive indices. The rock is one of the two most silicic members (SiO_2 = 58.2 %) analysed from the granulite belt.

US 15. Pyroxene granulite, near Biha, about 17.5 km NW of Matta. Composed of Plg, Cpx, Ore, Qz, Kf, Ap, Hbl, and Bio. Plg (An$_{50-52}$), and Kf(Or$_{80}$Plg$_{20}$) compositions determined from RIs.

SH 16. Foliated hypersthene gabbro (pyroxene granulite), 13 km N of Madyan, Swat Kohistan. Pyroxene analyses taken from Jan and Kempe (1973, Table 2). Analysts: C.J. Elliot, V.K. Din, and A.J. Easton.

US 24. Websteritic granulite in amphibolites, 4 km NE of Parao bridge, Upper Swat. Contains 62.3 % Cpx, 18.7 % Opx, 14.0 % tschermakitic Hbl ($Mg/(Mg+Fe^{2+}+Fe^{3+}+Mn) = 0.74$), and 5.0 % of secondary talc, Am, Ore.

SI 182. Pyroxene granulite, 3.5 km NNE of Seo in a road-cut, Indus Kohistan. Comprised of 71.0 % Plg (An$_{50}$), 13.0 % Opx, 11.8 % Cpx, 2.1 % Ore, and 2.0 % Am, Ap, etc.

SI 188a. Pyroxene granulite from a road cut near the bend in the Indus river, Banda Sazin. Mode: 72.0 % Plg (An$_{50}$), 12.3 % Opx, 9.6 % Cpx, 4.8 % Ore, and 1.4 % Ap, Qz, and secondary Am, Epi, Bio.

SI 188b. Pyroxene granulite intimately associated with but finer grained than 188a. Composed of 71.7 % Plg (An$_{51}$), 14.3 % Opx, 7.8 % Cpx, 5.3 Ore, and 1.0 % Ap, Bio, Am.

SI 192. Pyroxene granulite along road about 1.5 km E of the Harban Stream, Indus Valley. Contains 75.8 % Plg (An$_{55}$), 13.7 % Opx, 7.2 % Cpx, 1.1 % Hbl ($Mg/(Mg+Fe^{2+}+Fe^{3+}+Mn) = 0.61$, $Ca/(Ca+Na+K) = 0.76$), 1.0 % Qz, 0.6 % Bio ($Mg/(Mg+Fe^{2+}Fe^{3+}+Mn) = 0.61$), 0.5 % Ore, and 0.1 % Ap.

SI 193. Pyroxene granulite, along road 2 km W of Thor stream, Indus Valley. Composed of 69.9 % Plg (An$_{45}$), 7.3 % Opx, 8.1 % Cpx, 10.0 % Qz, 3.0 % Ore, 0.5 % Ap, and 1.1 % Bio, Am, etc. (Bio $Mg/(Mg+Fe^{2+}+Fe^{3+}+Mn) = 0.53$).

SI 197. Pyroxene granulite, along road 6 km E of the Thor stream. Consists of 63.1 % Plg (An$_{49}$), 11.6 % Opx, 8.3 % Cpx, 2.8 % Ore, 1.4 % Qz, 8.8 % Am, and 4.0 % Bio, Ap, Epi, Chl, etc. Hydrous minerals mostly retrograde.

SI 217. Pyroxene granulite 3.5 km N of Seo, on the western side of the Indus. Contains about 62 % Plg (An$_{55}$), 21 % Hbl ($Ca/(Ca+Na+K) = 0.70$), 8 % Cpx, 7 % Opx, 1 % Ore, 1 % garnet, and traces of Ap and (?) Qz. The pyroxenes have not been analysed but the compositions of Plg and Hbl have been used to estimate T.

ST 221. Pyroxene granulite along Knadia river, 4 km N of Richa. Contains Plg, Hbl, Cpx, Opx, Qz, Ore, Bio, and Ap. The Hbl and Bio are retrograde products developed under amphibolite facies conditions.

SK 322. Pyroxene granulite 1 km N of Madyan, east of river, Swat Kohistan. Mode: 58.5 % Plg (An$_{53}$), 22.1 % Opx, 16.0 % Cpx, 2.3 % Ore, 0.7 % Qz, 0.3 Am, 0.1 % Ap.

SI 324. Metasedimentary horizon in the amphibolites, Patan Stream, about 6 km NNW of Patan, Indus Valley. Contains Qz, Plg, Gar (? grossular), calcite, Cpx, and microcline.

SK 340. Pyroxenite member of the granulites near Pardesha, 3½ km NW of Madyan, Swat Kohistan. Consists of 75.4 % Opx, 8.7 % Plg, 8.4 % Cpx, 7.2 % of mostly retrograde Am, 0.3 % Ore, Opx (En$_{78}$) composition optically determined.

SK 354. Clinopyroxenite "layer" (containing Opx, Hbl, Plg, and Ore also) in the granulites N of Shahgram, 1 km NW of Madyan. Composition of the clinopyroxene ($Mg_{35}Fe_{17}Ca_{48}$) optically determined.

SK 422. Pyroxene granulite along path SW of Arin, 4 km WWN of Bahrain, Swat Kohistan. Contains 69.9 % Plg (An$_{48}$), 10.7 % Opx, 7.0 % Cpx, 7.4 % Qz, 2.3 % Ore, 1.9 % Bio, 0.4 % Ap, and 0.4 % Am, etc.

SK 448. Quartz diorite, in road-cut 2 km S of Kalam, Swat Kohistan. Composed of Plg, Bio, Opx, Cpx, Qz, Hbl, and Ore. (Bio $Mg/(Mg+Fe^{2+}+Fe^{3+}+Mn) = 0.57$).

SK 462. Epidote-biotite amphibolite from the isolated outcrop between road and river, opposite Asrit village, Swat Kohistan. Cpx composition ($Mg_{35}Fe_{15.5}Ca_{49.5}$) optically determined.

SK 528. Pyroxene granulite, in Bishigram (Madyan) stream, 1/2 km E of Karamai, Swat Kohistan. Contains 64.3 % Plg (An$_{50}$), 19 % Opx, 9.6 % Cpx, 4.9 % Qz, 1.4 % Ore, 0.4 % Am, 0.3 % Ap, and 0.2 % Bio.

SK 592. Pyroxene granulite, along southern Mankial Stream, 3/4 km S of Sur Gat, Swat Kohistan. Consists of 62.3 % Plg, 6.8 % Opx, 3.2 % Cpx, 18.0 % Qz, 4.6 % Bio, 3.2 % Am, and 1.9 % Ore, Ap, Epi, Chl. Hydrous minerals retrograde.

SK 600. Pyroxene granulite about 2 km SE of Mankial streams confluence NE of Kafir Banda. Composed of 71.1 % Plg (An$_{54}$), 16.1 % Opx, 9.9 % Cps, 1.4 % Ore, 1.0 % Qz, and 0.5 % Am, Bio, Ap.

SK 615. Pyroxene granulite, along stream, 2.5 km WWN of Asrit and SW of Shar Banda, Swat Kohistan. Contains 74.7 % Plg (An$_{46}$), 10.3 % Opx, 4.3 % Cpx, 6.6 % QZ, 2.2 % Ore, 1.4 % Bio, and 0.4 % Ap.

SK 647. Dioritic rock (? retrograde granulite) along foot path in ridge 2.8 km W of Dabna Swat Kohistan. Contains Plg (An$_{46}$), Hbl, Opx, Cpx, Bio, Qz, Ore, Ap, Hbl and Bio poikiloblastic and contain relics of pyroxene. Opx relics generally altered to chlorite, talc, and magnetite.