

tained from reciprocal or direct lattice constant variance-covariance matrices. The subscripts m and n assume values 2, 3 or 4 and 6 according to whether δ is a function of the edges or of the unit cell angles. Furthermore:

$$\delta\alpha/\alpha = (1/2A)(\delta B/\alpha B) \pm 1/\sqrt{(B^2-4AC)} [(B/2A)(\delta B/\alpha B) - (\delta C/\alpha C)] \quad (11)$$

$$\delta B/\alpha B = (\delta y_1 - \delta y_2) / (y_1 - y_3 - y_2 - y_1)$$

$$\delta C/\alpha C = \delta y_1^2 (y_2 - y_3) + \delta y_2^2 (y_4 - y_3) - \delta y_1 \delta y_2 (y_1 - y_3 - y_2 - y_1)$$

$$\delta\alpha/\alpha = (1/2A)(\delta B/\alpha B) \pm 1/\sqrt{(B^2-4AC)} [(B/2A)(\delta B/\alpha B) - (\delta C/\alpha C)] \quad (12)$$

$$\delta B/\alpha B = (\delta y_1 - \delta y_2) / (y_1 - y_3 - y_2 - y_1)$$

$$\delta C/\alpha C = \delta y_1^2 (x_2 - x_1) - \delta y_2^2 (x_1 - x_2) + \delta y_1 \delta y_2 (x_1 - x_2 - x_2 - x_1)$$

Equations (11) and (12) must be used with the positive sign or the negative sign according to whether the value of δ accepted has been obtained from (6) with the positive or negative square root of the discriminant.

Variance and covariance terms required to solve equation (10) are usually available on the output of the lattice constant refinement programs. Incidentally, for correct utilization of (10) it should be borne in mind that, as far as unit cell angles are concerned, computer programs usually furnish variance and covariance values in rad^2 and standard deviation values, i.e. the square root of the variances, in degree.

If on applying (8) one assumes that the errors are uncorrelated, i.e. $\text{cov}(V_{mm}, V_{nn}) = 0$, then

$$V_{\alpha} = \sqrt{\delta\alpha^2}, \quad V_{\beta} = \sqrt{\delta\beta^2}, \quad \text{and (10) reduces to } \delta\alpha^2/\alpha^2 = (\delta\alpha/\alpha)^2 V_{mm}^2 + (\delta\beta/\beta)^2 V_{nn}^2 \quad (13)$$

The values of the standard deviation given by (10) are generally slightly greater than those furnished by (13), but in both cases are smaller than those obtained graphically.

Errors arise in α and β . The standard deviation $\delta\alpha(\alpha)$ or $\delta\beta(\beta)$ can be calculated in a similar way to that of δ by means of the Law of Propagation of Errors:

$$\delta\alpha^2(\alpha) = (\delta\alpha/\alpha)^2 V_{mm}^2 + (\delta\beta/\beta)^2 V_{nn}^2 + 2(\delta\alpha/\alpha)(\delta\beta/\beta) V_{mn} \quad (14)$$

where V_{mm} , V_{nn} , V_{mn} have been previously defined. In this case, the subscripts m and n of course assume values 2, 3 or 4 in the reciprocal lattice constant variance-covariance matrix. Moreover:

$$\delta\alpha/\alpha = (1/2A)(\delta B/\alpha B) \pm 1/\sqrt{(B^2-4AC)} [(B/2A)(\delta B/\alpha B) - (\delta C/\alpha C)] \quad (15)$$

$$\delta B/\alpha B = (\delta\alpha_1 - \delta\alpha_2) / (y_1 - y_3 - y_2 - y_1)$$

$$\delta C/\alpha C = \delta\alpha_1^2 (y_3 - y_1) + \delta\alpha_2^2 (y_4 - y_1) - \delta\alpha_1 \delta\alpha_2 (y_1 - y_3 - y_2 - y_1)$$

$$\delta\alpha/\alpha = (1/2A)(\delta B/\alpha B) \pm 1/\sqrt{(B^2-4AC)} [(B/2A)(\delta B/\alpha B) - (\delta C/\alpha C)] \quad (16)$$

$$\delta B/\alpha B = (\delta\alpha_1 - \delta\alpha_2) / (y_1 - y_3 - y_2 - y_1)$$

$$\delta C/\alpha C = \delta\alpha_1^2 (x_3 - x_1) - \delta\alpha_2^2 (x_1 - x_2) + \delta\alpha_1 \delta\alpha_2 (x_1 - x_2 - x_2 - x_1)$$

When applying (14), it should be noted that double signs must be used in expressions (15) and (16) in a similar way to those used in calculating the errors in δ . If one assumes that the errors are uncorrelated the cross product of equation (14) becomes zero.

Acknowledgments. I wish to thank Prof. Joseph V. Smith for his encouragement and advice, and Prof. Giuseppe Schiavino for valuable suggestions.

Financial support was provided by C.N.R. (Italian National Research Council). All computations were carried out on an Univac 1106 computer at the Milan University Computer Center.

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J.F.W.Bowles : History of Fe₃O₄-FeTiO₃ grains (App.)

An estimation of the probable errors of the method of tracing the cooling history of complex magnetite-ilmenite grains and a discussion of the results produced by using different methods of treatment of the minor elements contained in these minerals when using the Buddington and Lindsley (1964) geothermometer.

(Appendix to "A method of tracing the temperature and oxygen-fugacity histories of complex magnetite-ilmenite grains", this vol. pp 103-109)

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Estimation of errors in the method of calculation used by Bowles (1977)

The formation of ilmenite from titanomagnetite frequently shows separate generations of exsolved ilmenite. Bowles (1977) used microprobe analyses of host magnetite and two generations of ilmenite with an estimate of the volume of each phase to derive compositions representing different stages in the cooling history of a complex magnetite ilmenite grain. Use of the Buddington and Lindsley (1964) geothermometer provided two points in the temperature and oxygen-fugacity history of the specimen. In this appendix a comparison of the errors of measurement is used to determine the likely errors in the result, to check the validity of that result.

The quantity of TiO₂ in the titanomagnetite is directly related to the quantity of ulvöspinel that is calculated. One standard deviation of the microprobe X-ray counts leads to an estimated error of $\pm 0.12\%$ TiO₂ on the measured value of 8.07% TiO₂. This error encompasses the X-ray statistical variation and the mineralogical variation between 20 measured localities on 5 adjacent magnetites. Only minor amounts of TiO₂ are subtracted when the minor elements are expelled to give $23.7 \pm 0.35\%$ ulvöspinel. Since the magnetite-ulvöspinel contours on the Buddington and Lindsley (1964) diagram are the most steeply inclined they have the greatest influence on the horizontal, temperature axis. At the condition of equilibrium of the magnetite with the ilmenite lamellae a variation of 10% ulvöspinel results in a temperature variation of 25°C, hence the calculated error of 0.35% ulvöspinel gives a temperature error of $\pm 0.88^\circ\text{C}$ and a value

of $\pm 1^\circ\text{C}$ is adopted here.

The quantity of total FeO contained in the ilmenite lamellae is 41.19% with an error of $\pm 0.33\%$ estimated from one standard deviation of the 20 microprobe measurements on ilmenite lamellae within the five adjacent magnetite grains. Only a small proportion of the measured Fe forms the 3.02% hematite but since a subtraction is involved the whole of the error has been loaded onto the hematite. The less steeply inclined ilmenite-hematite lines of the Buddington and Lindsley diagram indicate a greater influence on oxygen fugacity and at the lower temperature equilibration 2% hematite represents a change of about 1 log fO₂ in oxygen fugacity. Thus an error of 0.33% hematite represents an error of 0.17 log fO₂ oxygen fugacity and a value of ± 0.2 is taken here.

At the condition of equilibrium of the granular ilmenite the influence of errors in the microprobe results for the granular ilmenite is less, since the lines on the Buddington and Lindsley diagram are spaced more closely and reasoning similar to the above leads to -11.5 ± 0.04 log fO₂ for the oxygen fugacity measurement. In the determination of the earlier magnetite composition, errors introduced by other techniques must be considered. The change in cell size of the magnetite with composition is small, so the effect of errors in composition on cell size may be neglected. However the influence of the less precise point counting technique must be considered here. The same five adjacent magnetite grains for which microprobe results were obtained, were covered by point counting on equally spaced linear traverses and 1019 points were counted. The five grains in different orientations show wide variation in the width of the ilmenite lamellae at their intersection with the polished surface. However, since for each grain the relative orientation of the magnetite to the ilmenite is preserved, the measured value of the ratio of area is little affected by changes in orientation. The proportion of ilmenite lamellae determined in this way is $16.1 \pm 2.03\%$ with the error determined at the 95% confidence limit. This figure is used (Bowles, 1977) to determine that the number of atoms of Fe and Ti within the ilmenite is $(6.12 \pm 0.76) \times 10^{21}$ and $(33.60 \pm 0.84) \times 10^{21}$ within the magnetite. A total of $(39.72 \pm 1.14) \times 10^{21}$ atoms, of which $(5.61 \pm 0.16) \times 10^{21}$ are Ti, are calculated for the magnetite producing a Ti/(Ti + Fe) ratio of 0.141 ± 0.004 and leading to a molecule containing $42.3 \pm 1.12\%$ ulvöspinel. At the conditions of equilibration of the granular ilmenite a change of 10% ulvöspinel corresponds to a 29°C change in temperature and leads to a temperature of $930 \pm 3.2^\circ\text{C}$. Because the lines on the Buddington and Lindsley diagram intersect at a low angle, an error on one set of lines influences estimates of error on the other axis

and the error in oxygen fugacity due to the microprobe results is masked by the larger error (0.1 log f_{O_2}) measured from the diagram.

The 16.1% of the second generation ilmenite lamellae changes the magnetite composition by 18.6 Mol %. The third generation of ilmenite lamellae occupy less than 0.1% of the specimen. These lamellae can only change the molecular percentage of the ulvöspinel in the magnetite by less than one third of the 0.35% error of the microprobe results, so that they have not been considered in the calculation.

Mineralogical measurements can clearly be placed on the Buddington and Lindsley diagram with a good precision but the accuracy of the whole set of results is governed by the error limits of $\pm 30^\circ\text{C}$ and $\pm 1 \log f_{O_2}$ laid down by Buddington and Lindsley for their experimental work.

Discussion of the method of treatment of minor elements used by Bowles (1976, 1977)

Complete analyses of naturally occurring iron-titanium oxides often show appreciable concentrations of other elements. A number of methods of dealing with these impurities are in common use. One method is proposed by Chevallier, in Vincent, Wright, Chevallier, and Mathieu (1957), whilst another method, due to D. R. Wones, is described by Buddington and Lindsley (1974). The methods appear similar in that divalent elements are gathered together as R_0 , trivalent elements appear as R_2O_3 , and tetravalent form TiO_2 . However, the use of these groupings by various authors is very different. In the method of Chevallier, magnetite and magnetite-like molecules (eg MgO, Al_2O_3) are grouped together and taken as magnetite. TiO_2 and SiO_2 are equated to form ulvöspinel by the addition of 2 FeO, and the ulvöspinel is taken to include ulvöspinel-like molecules such as 2 MgO, TiO_2 , where some of the TiO_2 in the entire group is in fact SiO_2 . Remaining R_0 and TiO_2 are associated to form an ilmenite suite of minerals. Molecular proportions of magnetite, ulvöspinel and ilmenite are calculated and thus contain the impurities distributed between them in an arbitrary manner. A similar treatment has been described by Carmichael (1967a). The method of Buddington and Lindsley groups and discards R_0, R_2O_3 and $2R_0, SiO_2$ but in this method R_0 and R_2O_3 do not contain iron oxide. Either R_0 or R_2O_3 are left, and these are grouped with small amounts of iron oxide and TiO_2 as FeO, R_2O_3 or as equal amounts of R_0, Fe_2O_3 and $2R_0, TiO_2$, and discarded from the system. Remaining iron and titanium oxides are grouped so as to lie on the appropriate solid solution series.

Anderson (1968) has proposed a variation on this method in which the treatment of the minor elements in the magnetite is changed, following the work of Speidel (1967), so that $2R_0, TiO_2$ is preferred to R_0, Fe_2O_3 , and R_0, R_2O_3 contains a proportion of FeO as R_0 . A more comprehensive discussion of the reasons why Anderson made this change is contained in a later paper by Speidel (1970).

Buddington and Lindsley suggest that, if the impurities have not been measured, then the iron-titanium oxides can be rounded up to 100%. This method is occasionally used (Duchesne 1972) where the impurities have been measured and then discarded. However, this method can only be applied where Fe^{2+} and Fe^{3+} have been determined, and the major phases can be computed without recourse to the minor elements. If this ratio has not been determined, as with microprobe analyses, it is necessary to associate the minor elements with small amounts of the major elements before the major phases can be assembled by distributing total Fe between Fe^{2+} and Fe^{3+} to form ilmenite and hematite or ulvöspinel and magnetite. If the results are to be used in conjunction with the Buddington and Lindsley geothermometer, it is essential to decide which method should be used. Since the experimental system of Buddington and Lindsley involved only the pure $FeO - TiO_2 - Fe_2O_3$ system, the methods of Chevallier and Carmichael are inappropriate here. The methods of Chevallier and of Buddington and Lindsley require that the Fe^{2+} to Fe^{3+} ratio is known and as such are inappropriate for microprobe results. The method of Wones acknowledges the minor elements, yet gives the pure system as the final result. This method, as modified by Anderson, appears to be the most satisfactory and is the method selected here and used elsewhere (Bowles 1976, 1977). A specimen previously described (Bowles 1977) is used in Table I to illustrate the difference in the results produced by the various methods.

By adopting different methods of calculation of the distribution of trace elements, significantly different results may be obtained. This is illustrated by the work of Elsdon (1972) in which he has made a comparison between iron-titanium oxides of the Kap Edvard Holm layered gabbro calculated according to Anderson, and the same phases from Skaergaard as calculated by Buddington and Lindsley, from the analyses of Vincent and Phillips (1954). Table II shows one of Elsdon's analyses (No 2) recalculated by these methods and it is seen that the method of Anderson gives a result some 80°C lower and with an oxygen fugacity about 1.5 to 2.5 more negative on the log f_{O_2} scale. Elsdon's comparison showed that the Kap Edvard Holm series equilibrated at a temperature 200 to 300°C lower than Skaergaard but it is

seen that a proportion of this temperature difference is caused by the different method of calculation of the results. However the change in the oxygen fugacity caused by the change in calculation is twice as large and in the opposite direction to the comparison made by Elsdon. As it happens this does not detract from Elsdon's argument, since the change brought about by changing the method of calculation has the effect of sliding the results along the buffered cooling curve.

It is therefore worth noting that analytical results can, in some cases, be affected quite strongly by the method of calculation of the minor elements, and if comparisons are to be made they should clearly be between results calculated in the same manner. The method due to Wones (Buddington and Lindsley, 1964) as modified by Anderson (1968) is preferred for calculations based on microprobe results destined for use on the Buddington and Lindsley geothermometer. Analyses should contain minor elements which are used to subtract a proportion of the Fe and Ti, to leave a result which may be compared with the pure experimental system of Buddington and Lindsley. Comparison of the remaining Fe and Ti with appropriate formulae enables the total Fe to be allocated to Fe^{2+} and Fe^{3+} .

Acknowledgements: This work has benefitted from the encouragement of Dr. M. K. Wells and financial assistance from NERC and University College London.

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TABLE I A comparison of the results obtained for the molecular composition of a single specimen of magnetite and ilmenite lamellae after the minor element distribution has been calculated using several published methods.

The analysis of these phases was given by Bowles (1977 p. 107).

Method of calculation of minor elements	Chevallier	Wones	Buddington & Lindsley	Carmichael	Anderson
Reference	Vincent <u>et al.</u> (1957)	Buddington <u>et al.</u> (1964)	Buddington <u>et al.</u> (1964)	Carmichael (1967a)	Anderson (1968)
Method adopted in papers by:	Vincent <u>et al.</u> (1957)		Buddington (1964)	Carmichael (1967a) Elsdon (1972)	Anderson (1968)
			Duchesne (1972)	Carmichael (1967b) Nicholls (1971)	Czamanske (1972)
				<u>et al.</u> (1972).	
Ilmenite %	96.0*	96.6	87.6	96.0*	96.6
Hematite %	4.0†	3.4	12.4**	4.0*	3.4
Magnetite %	72.9	74.0	73.5	72.9	76.3
Ulvöspinel %	27.1	26.0	26.5	27.1	23.7
Temperature (°C)		678		675	662
Log f_{O_2}		-18.2		-18.5	-19.0

* R_0, R_2O_3 † R_2O_3 ** TiO_2

(Table II overleaf)

TABLE III. Comparison of calculations of results given by Eisdon (1972) for his specimen No. 2.

Method of calculation of minor elements	Wones	Buddington & Lindsley	Anderson	Eisdon's Calculations
Ilmenite	87.7	87.7	87.7	87.7
Hematite (Mol %)	12.3	12.3	12.3	12.3
Magnetite	75.2	73.4	79.3*	81.2
Ulvöspinel (Mol %)	24.8	26.6	20.7	18.8
Temperature (°C)	771	786	706	700
Log f_{O_2}	-14.3	-14.0	-15.8	-16.5

* The difference in results for magnetite - ulvöspinel are probably due to difference in the Mg/Fe allocated in RO to form $RO_2R_2O_3$.

S. Sinha Roy: Metamorphism in the Himalayas

METAMORPHISM IN A HIMALAYAN THRUST ZONE

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The Proterozoic Daling Formation of the Sikkim to Darjeeling area of the Himalayas occurs as a tectonostratigraphic allochthonous unit above a parautochthonous Upper Palaeozoic (Lower Gondwana) sequence (Ganewar, 1964, Sinha Roy, 1973a). The Lower Gondwana rocks are separated in isolated tectonic windows in the Sikkim Himalayas and are separated from the Daling and associated rocks by well defined thrust zones (Ghosh, 1956). The thrusts are considered to be deep-rooted because pre-Daling basement rocks occur along them as tectonic slices. The Daling rocks in the lower zones are meta-greywackes, psammites and amplitites and basic igneous rocks. They show evidence of at least three phases of deformation and metamorphism (Sinha Roy, 1973b). Age data indicate that the main (Barrovian) metamorphism is linked with the Tertiary Himalayan orogeny (Crawford, 1974) The Barrovian zones are inverted (Ray, 1967; Ganewar, 1964a), and in the vicinity of the thrust zone under consideration, are of lower greenschist facies.

At the base of these greenschist facies rocks within a meta-greywacke unit of the thrust zone, there are moderate to high-pressure assemblages which pass into the more extensive greenschists and are categorically developed for 200 to 250 m above the thrust. **Nodules to high-pressure metamorphic assemblages.** The porphyroclastic components of the meta-greywackes are quartz, albite and oligoclase. These are highly deformed into lenses with pressure fringes in the foliation, the latter being a mylonitic fabric. The groundmass comprises recrystallized fine polygonal quartz and feldspar and sericite. All or some of the following minerals may occur in single specimens. Stilpnomelane as some fibrous aggregates replacing plagioclase and as stumpy grains enclosed within the fold of chlorite and sericite and parallel to the foliation. Lawsonite (2 to 3%) occurs as either tabular grains (0.02 mm long), as small clusters replacing plagioclase, and in the fold of chlorite and sericite along the foliation. Near the thrust, phengitic mica is present in minor amounts as stumpy flakes orientated sub-parallel to the foliation and replaced along the cleavages by stilpnomelane. Here, the lawsonite-bearing assemblages contain aragonite which occurs as a thin film parallel to the foliation, as aphelesite grains replacing plagioclase and as subhedral grains enclosed with quartz and lawsonite. Pumpellyite is first met about 150 m above the thrust where it occurs as a fibrous felt orientated parallel to the foliation and associated with small grains of lawsonite. It increases appreciably in amount (up to 20%) away from the thrust and some 250 m above the thrust it defines polygonal arcs of microfolds and also occurs as slender needles in radiating clusters. Stilpnomelane, orientated parallel to the axial planes of such microfolds, often replaces the pumpellyite while subhedral epidote replaces both the stilpnomelane and the pumpellyite in the cores of the microfolds. As the Barrovian biotite zone is approached lawsonite is also replaced by epidote and biotite grow from stilpnomelane, pumpellyite, epidote and chlorite. Accessories in these assemblages are zircon, apatite, sphene, tourmaline and opaque. Thus the assemblages at or near the thrust are quartz-plagioclase-epidote-sericite-chlorite-tillomelane-lawsonite-aragonite and quartz-plagioclase-sericite-chlorite-phengite-stilpnomelane. Away from the thrust and with passage into greenschist-facies the assemblages are, quartz-plagioclase-chlorite-sericite-epidote, quartz-plagioclase-chlorite-pumpellyite-tillomelane-epidote and quartz-plagioclase-chlorite-sericite-lawsonite-epidote. In the biotite zone of the adjacent greenschist facies there are quartz-plagioclase-chlorite-muscovite-biotite-epidote, and quartz-plagioclase-chlorite-sericite-biotite.

The textural evidence suggests that the assemblage lawsonite-quartz-sericite was stable and this indicates a transition to high-pressure conditions (Glenisley et al., 1976). The presence of oligoclase is incompatible with relatively high-pressure conditions (Taylor and Coleman, 1968) but here the mineral is a relict elastic phase. The rocks may therefore be considered to belong to the lawsonite-biotite facies (Winkler, 1967) or high-pressure intermediate facies series (Miyashiro, 1961). Crawford and Fyfe (1969) and Misch (1968) have demonstrated that stable coexistence of lawsonite and quartz requires 4 to 5 kb and the minimum pressure required has been shown experimentally to be 3.2 kb at about 250° and 3 kb at 210° C where $P_{H_2O} = P_{total}$ (Liu, 1971). Zen (1974) argues, contrary to Miyashiro (1961) that pumpellyite-bearing assemblages may not indicate high pressure and low temperature metamorphism. He, however, indicates that pumpellyite of specific composition could be stable at pressures of from 4 to 7 kb, and that although higher pressures seem to favour pumpellyite, its stability relations at low pressure and in heterogeneous equilibria are uncertain.

The metabasic rocks associated with the meta-greywackes contain an amphibole which is considered to be intermediate between glaucophane and actinolite (N. L. Dobretsov, personal communication; Sinha Roy, 1975). Other minerals in the meta-basites at or near the thrust are pumpellyite, stilpnomelane and rarely, lawsonite. In the higher structural levels,

actinolite replaces the intermediate amphibole and the pumpellyite-actinolite-bearing assemblage in the metabasic rocks and the pumpellyite-chlorite-bearing assemblage in the meta-greywackes would indicate that these rocks belong to a higher grade than the prehnite-pumpellyite facies (Coombe, 1960). The disappearance of the pumpellyite-bearing assemblage in the greenschist facies is to be expected (Coombe et al., 1970). At temperatures of 200 to 300° C the pressure necessary for the formation of aragonite from calcite is about 5 to 8 kb (Crawford and Fyfe, 1964; Roettcher and Wyllie, 1967). If the aragonite is stable in the lawsonite-bearing assemblage, then the pressure indicated would be around 6 kb if it is assumed that the temperature is about 250° C. This is within the stability field of lawsonite (cf. Newton and Kennedy, 1963), and is lower than the temperature probably reached in the adjacent greenschist facies (Roettcher and Wyllie, 1967; Johannes and Puhar, 1971). From these considerations, it may be suggested that the rocks in the thrust zones have undergone a moderate- to high-pressure and relatively low temperature metamorphism.

Discussion. It is generally held that rocks containing high-pressure assemblages are restricted to tectonic belts that are considered to be former continental margins, representing compressional plate boundaries. In the Himalayan tectogenes, the ideal location for such rocks would be the Indus-Tsangpo suture zone. The present thrust zone is about 200 km south of this suture and is close to the frontal thrust belt. It is therefore difficult to link this metamorphism with continental collision between India and Tibet. The rocks affected are Proterozoic and, if the metamorphism is Tertiary then the time relationships are unusual for in other similar belts the stratigraphic age of the rocks is not much greater than the age of the metamorphism (Sinha Roy, 1972). It is difficult to decide whether the moderate- to high-pressure metamorphism is a Precambrian event which is partially or totally overprinted by Tertiary Barrovian metamorphism, or whether the Tertiary metamorphism is plurifacial (Sinha Roy, 1975). While relicts of an earlier high-pressure, low-temperature metamorphism within a later Barrovian metamorphism are considered to be a possibility in the Appalachian metamorphic belt (Zen, 1974), the former is related to the regional thrust-faulting in the plurifacial metamorphic environment in the Franciscan belt (Ghent, 1965; Blake et al., 1969). Because of the proximity of the present assemblages to the thrust the latter possibility is to be preferred and this is supported by the fact that the minerals are coeval with the thrusting and grow in and define the mylonitic foliation which in the greenschist facies rocks is an axial planar foliation to folds developed during the same deformation.

The reworked Proterozoic rocks in the inner tectonic belt, including the window zones of the Himalayas, are thought to have been derived through thrusting of a 'microcontinent' cover sequence, which lay between the land masses of India and Tibet (Sinha Roy, 1976a). The deep-rooted thrusts such as the present example, which are located to the south of the central crystalline axis and hence much to the south of the Indus-Tsangpo suture, represent a Tertiary collisional interaction zone between the microcontinent and the Indian continent (Sinha Roy, 1976b). If as suggested, the moderate- to high-pressure metamorphism is temporally related to the Barrovian metamorphism, then the former might be the result of such an interaction.

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D.R.Mason: Zoned amphibole from Papua

M19

Zoned Amphibole in the Yirri Intrusive Complex, Manus Island, Papua New Guinea.

The Yirri Intrusive Complex (Jaques and Webb, 1975) of mid-Miocene age occupies about 200 sq. km in the central and southern parts of Manus Island, about 400 km N of the Papua New Guinea mainland. The island forms the NW extremity of the Tertiary Outer Melanesian Arc.

High-K dioritic rocks (Gulson et al., 1972) constitute the bulk of the Complex, and contain a typical calc-alkaline assemblage of plagioclase, K-feldspar, quartz, amphibole, biotite, magnetite, and accessory sphene and apatite. The amphiboles of dioritic rocks have usually been reported as being of 'magnesian-hornblende' type in the terminology of Leake (1968). Chemical analyses of a high-K diorite and a low- to normal-K diorite from the Mount Kren area are given in Table 1.

Electron microprobe analyses (Ware, 1973; Reed and Ware, 1973) of discrete amphibole grains in these rocks reveal a wide range of composition (Table 1). In the high-K diorite, green amphibole is zoned from magnesio-hornblende in cores to actinolitic hornblende in rims. The zoning is not evident in thin section. In the low- to normal-K diorite, amphibole is zoned from ferroan pargasitic hornblende cores, through edenitic hornblende, to magnesio-hornblende rims. In thin section the pargasitic cores can be discerned as sharply-defined, central brown areas entirely surrounded by green amphibole ultimately forming discrete subhedral grains up to 2 mm long. Fig. 1a shows a positive correlation between Al^{IV} and total Fe, a relationship attributed to high f_{O_2} during crystallization by Czamanske and Wones (1973). Such an interpretation is particularly acceptable when it is noted that hydrothermal alteration and porphyry copper mineralization are present in the Complex (Jaques and Webb, 1975). The negative correlation for the pargasitic cores might imply that conditions of low f_{O_2} were operative at the time of formation of the core amphibole.

The pargasitic cores have compositions quite atypical of amphiboles from dioritic rocks. With about 43% SiO_2 , 3% TiO_2 , and 10% Al_2O_3 they are more characteristic of amphiboles from gabbroic rocks and of high-pressure, high-temperature amphiboles from experimental work on basaltic rock compositions (Table 1). It might be suggested that the dioritic rocks of the Complex have assimilated basaltic country rocks, but the intrusive rocks are texturally homogeneous and mafic xenoliths are rare or absent.

The preferred interpretation is that the brown pargasitic cores represent relic material from a partially melted mafic source (possibly in the base of the crust), and the green magnesio-hornblende mantles represent amphibole

Table 1: Analyses of diorites and their amphiboles, Yirri Intrusive Complex, Manus Island, Papua New Guinea.

	1	2	3	4	5	6	7	8
SiO_2	61.58	59.91	48.9	52.9	43.7	51.5	42.1	40.2
TiO_2	.48	.53	1.5	.4	3.0	.7	2.4	3.1
Al_2O_3	16.92	17.31	5.8	3.2	10.3	3.9	12.5	15.4
Fe_2O_3	2.51	2.99						
FeO	1.81	2.10	9.9*	9.5*	10.8*	9.3*	13.8*	9.8*
MnO	.04	.05	.3	.4	.2	.3	.3	-
MgO	2.72	2.99	17.3	17.8	15.2	18.1	12.5	12.1
CaO	5.18	6.63	11.4	12.3	11.6	11.9	12.1	11.8
Na_2O	4.60	4.75	1.5	.7	2.5	1.2	2.0	1.8
K_2O	2.49	.88	.5	.2	1.1	.3	.8	.3
P_2O_5	.29	.29						
S	.03	.03						
H_2O^+	.83	.95						
H_2O^-	.20	.19						
CO_2	.10	.09						
total	99.78	99.69	97.1	97.4	98.4	97.2	98.5	94.5

Explanation: * total Fe as FeO.

1: Specimen DRM110, high-K hornblende (biotite-clinopyroxene) diorite. 2: Specimen DRM112, low- to normal-K hornblende-quartz diorite. Both specimens from outcrops in Willie River, 2.5 km SW of Mount Kren. All elements by XRF except Na, FeO, and volatiles. Analysts: D.R. Mason and P.H. Beasley. Probe analyses. 3, 4: green core and green rim respectively of amphibole in DRM110. 5, 6: brown core and green rim of amphibole in DRM112. Core and rim analyses for DRM110 and DRM112 were selected from totals of 25 and 27 spot analyses respectively to display the maximum amphibole compositional variation obtained for each specimen. 7: average of 6 spot analyses of brown hornblende in DRM065, a gabbro with granulitic texture from the Western Highlands of Papua New Guinea (Mason, 1975). 8: Amphibole in basaltic composition at 10 kb and 920°C (Green and Ringwood, 1968). Note that the brown core of DRM112 is similar to amphibole from mafic rocks.

which has crystallized from the melt. This interpretation supports theories which consider most granitoids to be composed initially of a melt plus solid refractory material from the source (Pivinskii, 1968; Chappell and White, 1974).

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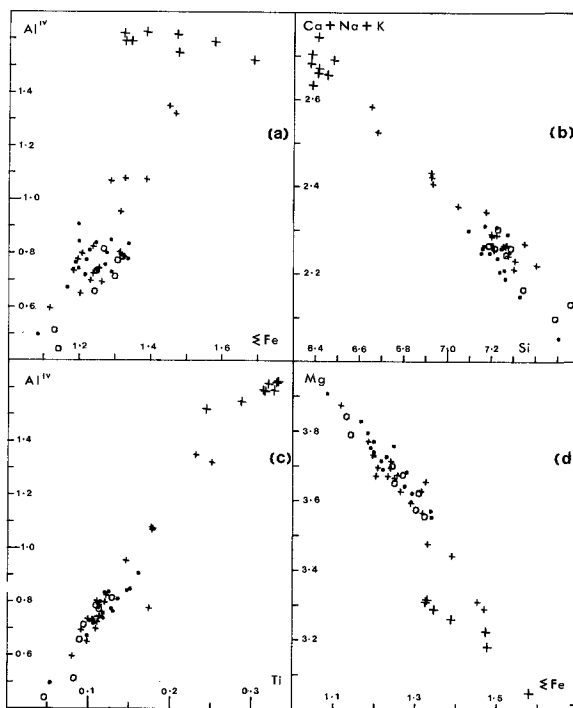


Fig. 1: Compositional variation in amphiboles, using data from structural formulae (cations per half unit cell). (a) Al^{IV} versus total Fe. (b) $Ca+Na+K$ versus Si. (c) Al^{IV} versus Ti. (d) Mg versus total Fe. Small filled circles: green cores from DRM110. Open circles: green rims from DRM110. Large crosses: brown cores from DRM112. Small crosses: green rims from DRM112.