

# Mineral parageneses and metamorphic reactions in metasedimentary enclaves from the Archaean Gneiss Complex of north-west India

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**ABSTRACT.** Three metasedimentary enclaves up to a kilometre in length of contrasting compositions within the polymetamorphic Banded Gneissic Complex ( $> 2580\text{Ma}$ ) have been studied for their mineral parageneses and metamorphic conditions. The largest enclave, consisting of kyanite-chloritoid-muscovite schist with quartz or corundum, and kyanite-fuchsite-corundum  $\pm$  diaspore, was metamorphosed at most under lower amphibolite conditions, and is thus not isofacial with the surrounding schists and gneisses (of the 'basement' complex) which reached sillimanite-grade metamorphism in the last orogenic cycle (Aravalli: 1650–950Ma Orogeny) in Rajasthan.

The second enclave is a calc-silicate rock which occurs as a small lens. The presence of two generations of wollastonite which formed during different metamorphic events in the calcite-quartz-grossularite-anorthite-clinopyroxene assemblage indicates polymetamorphism.

The third enclave is a metabasic rock which records a complete polymetamorphic history in discontinuous zones in garnet coexisting with hornblende-chlorite-plagioclase-quartz  $\pm$  epidote. To explain the garnet zoning a model involving partial resorption of early garnet during the initial recrystallization stage of superimposed regional metamorphism is preferred to the alternative based on a single prograde metamorphism and retrogression.

The mineralogy of the calc-silicate and metabasic enclaves gives a recrystallization temperature of  $c. 700^\circ\text{C}$  and a pressure in the range of 8–3 kbar during the second metamorphism.

**THE Banded Gneissic Complex—BGC** ( $> 2580\text{Ma}$ , Crawford, 1970)—is the oldest stratigraphic unit in Rajasthan in NW India (Heron, 1953) fig. 1). It is made up of polymetamorphic schists, gneisses, and migmatites. Besides being intruded by acid and basic rocks, the BGC contains metasedimentary inclusions of different ages, which are either synclinal cores of the overlying Aravalli System

rocks (2500–2100Ma) due to cofolding (Heron, 1953), or the restites formed as a consequence of anatexis suffered by the BGC (Sharma, in press) during the Aravalli Orogeny (1650–950Ma, Crawford, 1970). It is also possible that some enclaves are older than the complex itself, but they cannot be distinguished from the younger ones until radiometric dating combined with detailed petrological work is carried out on them.

This paper describes some of the assemblages from the metasedimentary enclaves around Chandera (formerly Chandra) in the Udaipur district of Central Rajasthan (see Sharma, 1983). These include: fuchsite-corundum-kyanite, chloritoid-kyanite-muscovite-quartz, garnet-hornblende-chlorite-plagioclase  $\pm$  epidote, and wollastonite-grossularite-calcite-anorthite-diopside-quartz. They are found as bands and lenses of different sizes (fig. 1) within the rocks of the BGC which in this region has been considered as granitized Aravallis by some workers (e.g. Naha *et al.*, 1967; Naha and Halyburton, 1974).

## Field occurrence

The main outcrop consisting of kyanite-chloritoid-muscovite  $\pm$  quartz schist and minor fuchsite-corundum-kyanite schist occupies a hill (1.2 km  $\times$  300 m) in the BGC. This outcrop was identified as talc-kyanite ( $\pm$  sillimanite) schist by S. K. Chatterjee (in Heron, 1953, p. 78). The kyanite schist and fuchsite-rich kyanite schist are foliated and lineated and have coarse porphyroblasts of kyanite. The kyanite-mica schist is flanked by quartzite and in turn by muscovite-quartz-biotite-K-feldspar-plagioclase  $\pm$  sillimanite schist/gneiss

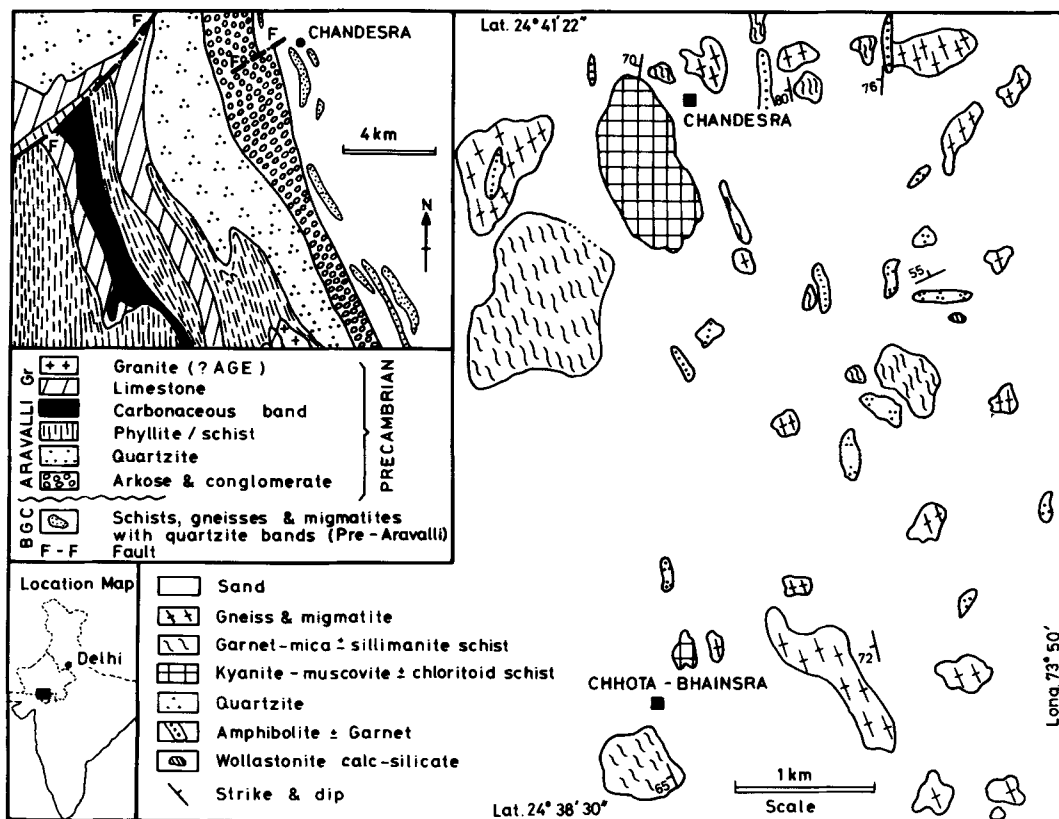


FIG. 1. Outcrop map around Chandesra, district Udaipur, Rajasthan, showing the occurrence of metasedimentary rocks in the Archaean Banded Gneissic Complex. Location map and regional geological map (after Heron, 1953) are shown as inset.

of the BGC (fig. 1). These enclave rocks dip 65–75° towards the SW and strike NNW conformable with the compositional layering in the surrounding pelitic schists, gneisses, and migmatites.

The garnet-chlorite-bearing amphibolite and the wollastonite-bearing calc-silicate rock occur as thin bands 5–10 m long and 1 m long respectively in the BGC schist. (fig. 1).

#### Petrography and mineral compositions

The following assemblages have been recorded in the enclaves:

##### A. Aluminous-rich compositions:

1. Ky-Fu-Co-Ru  $\pm$  Qz  $\pm$  Dias
2. Ky-Fu-Co-Ru
3. Mus-Co(-Qz)
4. Ky-Mus-Qz-Ru
5. Ky-Ctd-Mus-Ru-Co
6. Ky-Ctd-Mus-Qz-Ru
7. Ctd-Mus-Qz-Ru-Ilm  $\pm$  Stau

##### B. Calc-silicates:

1. Wo-An-Cc-Diop-Gar-Qz
2. Wo-Gar-An-Diop

##### C. Basic compositions

1. Gar-Hb-Chl-Plag  $\pm$  Ep  $\pm$  Qz
2. Gar-Hb-Chl-Plag  $\pm$  Ilm

Abbreviations: Cc = calcite; Chl = chlorite; Co = corundum; Ctd = chloritoid; Dias = diasporite; Diop = diopside; Ep = epidote; Fu = fuchsite; Gar = garnet; Hb = hornblende; Ilm = ilmenite; Ky = kyanite; Mus = muscovite; Qz = quartz; Ru = rutile; Stau = staurolite; An = anorthite.

*Fuchsite* mainly occurs as oriented greenish-coloured lepidoblasts which trend in the elongation direction of kyanite. However, small, diversely oriented flakes with less colouration are found along cracks in the kyanite. These two types are chemically similar except that the larger flakes have such a high Cr content (see Table I) that they are classified as fuchsite (Deer *et al.*, 1962); they are similar to the fuchsites from the Mashishimala deposit of Transvaal (Schreyer *et al.*, 1981). Both large and small flakes are free of the phengite component, and paragonite substitution ranges up to 13 mol. %.

*Muscovite* is ubiquitous in kyanite-bearing assemblages devoid of fuchsite. It defines the schistosity and records deformational features in the rock. The muscovite is often transected by chloritoid and kyanite and frequently co-folded with corundum and rutile (fig. 2b).

The composition of muscovite is similar, irrespective of its location or orientation in different assemblages (see Table I). Like fuchsite, the

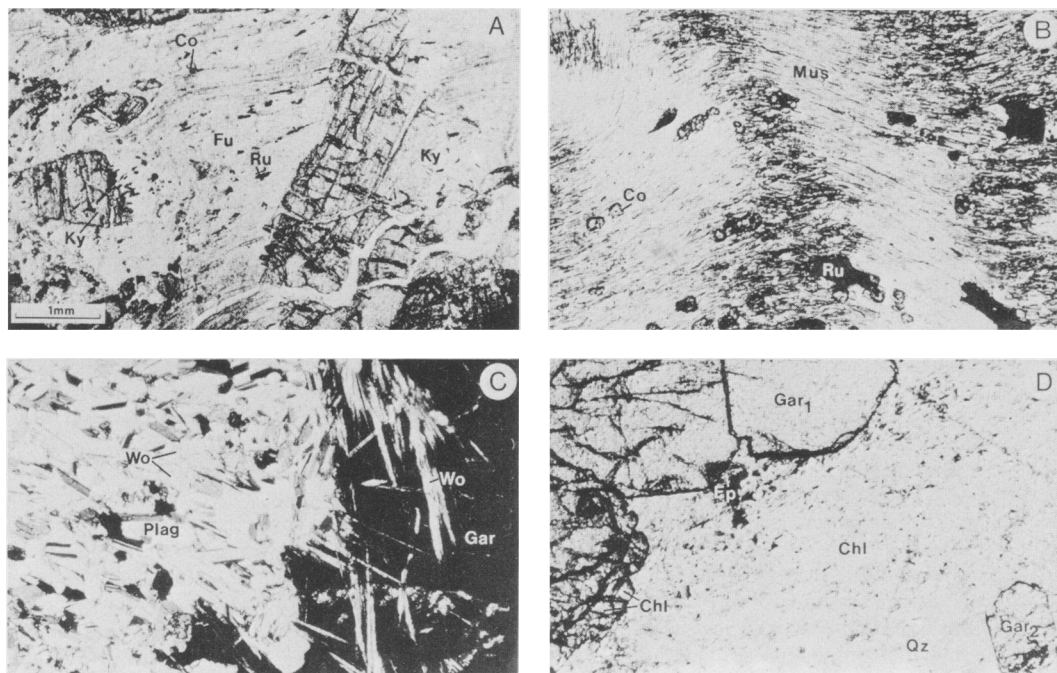


FIG. 2. Photomicrographs showing: (a) overgrowth of kyanite (Ky) on a fold outlined by rutile (dark grains) and corundum (Co). Observe that kyanite transects schistosity defined by fuchsite. The scale bar represents 1 mm. The other photomicrographs are all to this scale; (b) cofolding of corundum and muscovite. Notice concentration of corundum in fold hinges; (c) two generations of wollastonite (Wo). Early acicular type crystals occur exclusively within garnet (Gar) which also encloses calcite and diopside. The second generation wollastonite is seen as broad prisms which with anorthite (Plag) exhibit excellent dimensional orientation of groundmass; (d) small euhedral garnet (Gar<sub>2</sub>) of the second metamorphic event embedded in chlorite (Chl) matrix, while garnets of the first event are large, subhedral to anhedral, and often enclose hornblende, chlorite, and epidote (Ep) in their margins.

muscovite is almost devoid of the phengite component and is a solid solution of muscovite-paragonite with 12–14 mol. % paragonite.

Kyanite varies in modal amount from 20 to 55%. It is post-kinematic with respect to  $F_1$  because of its overgrowth on microfolds defined by rutile  $\pm$  corundum granules (fig. 2a). Kinks and cross fractures in this mineral result from the second deformation and indicate that kyanite crystallized between the  $F_1$  and  $F_2$  fold phases of the Aravalli orogeny (Sharma, 1977). Occasionally, kyanite is altered especially along its margins to muscovite. Kyanite from fuchsite-bearing schists has 0.3 to 0.5 wt. %  $\text{Cr}_2\text{O}_3$  (see Table I) and this favours the idea that the sedimentary protolith had a high  $\text{Cr}_2\text{O}_3$  content.

Chloritoid forms less than 10 vol. % and is mostly associated with kyanite. It is characteristically high in  $\text{Al}_2\text{O}_3$  and FeO and low in MgO and almost free of CaO, MnO, and alkalis. Since its structural formula relative to 12 oxygens shows  $\text{Fe} + \text{Mg} + \text{Mn} = 2$  with no deficiency in Al, the mineral must

be almost free of  $\text{Fe}^{3+}$ . One of the three brucite-type octahedral sites is occupied solely by Al and the corundum-type octahedral sites by 3Al. The two remaining brucite-type octahedral sites are filled mainly with  $\text{Fe} + \text{Mg}$ , almost without Mn.

These compositional characteristics point to the absence of oxidizing conditions. Further evidence of low  $f_{\text{O}_2}$  comes from the presence of rutile, ilmenite, and rare magnetite and the absence of hematite.

*Staurolite.* Only one grain has been recognized within chloritoid. It is partially altered to greenish chlorite and sericite, thus making its analysis difficult. However, in view of the straw-yellow colour and faint pleochroism in contrast to all associated phases, the mineral is reasonably identified as staurolite.

*Corundum* constitutes about 5–8 vol. % of the fuchsite-kyanite schist, but it occurs only in trace amounts in the quartz-free domains of the muscovite-kyanite + chloritoid paragenesis (fig. 2b). In the fuchsite-kyanite-schist it shows marginal

alteration to a yellowish-orange earthy mass (? diasporite). Its composition is almost pure  $\text{Al}_2\text{O}_3$ , containing only 0.2 wt. %  $\text{Cr}_2\text{O}_3$  and 0.25 wt. %  $\text{TiO}_2$ .

*Rutile* occurs as tiny yellowish to reddish-brown granules in all the aluminous-rich compositions. Microprobe analyses indicate as much as 0.86 wt. %  $\text{Cr}_2\text{O}_3$  (see Table I).

*Quartz* (6–10 vol. %) occurs as small crystals and more often in pockets of the muscovite-kyanite-bearing assemblages. It shows sieve texture owing to minute inclusions of muscovite.

Zircon and sometimes apatite are the accessory minerals.

The absence of albite, chlorite, garnet, and biotite is noteworthy, although the stable paragenesis of chloritoid-kyanite-muscovite-quartz indicates greenschist-facies conditions. It appears that low contents of  $\text{Na}_2\text{O}$  and near absence of  $\text{CaO}$  in the parent rocks, together with the incompatibility of chloritoid with albite in the greenschist facies (Hoschek, 1969), are responsible for the absence of albite. Also excess  $\text{Al}_2\text{O}_3$  and insufficient  $\text{K}_2\text{O}$  in the original sediments may have prevented the formation of biotite in these assemblages. Thus it seems that the compositional constraints of the parent rocks did not allow the formation of some expected phases. The relevant rock system is then definable in terms of  $\text{SiO}_2\text{--Al}_2\text{O}_3\text{--(FeMg)O--K}_2\text{O--(H}_2\text{O)}$ . Consequently the phases can be projected from  $\text{K}_2\text{O}$  on to the AFM plane (Thompson, 1957) from which it would appear that chloritoid is highly Fe-rich. This would, therefore, prevent almandine from being a stable phase with staurolite (now a relict phase) even if the reaction:  $\text{Ctd} + \text{Ky} = \text{Stau} + \text{Gar} + \text{H}_2\text{O}$  (Richardson, 1968) had occurred during prograde metamorphism. Moreover, it appears from chloritoid analyses (almost free of  $\text{MnO}$  and  $\text{CaO}$ ) that  $\text{MnO}$  and  $\text{CaO}$  were insufficient to stabilize garnet. The absence of chlorite is problematic in view of the fact that the rocks in the area have undergone retrogression in the greenschist facies and that chlorite is abundant in the basic assemblages. Probably the disappearance of chlorite is due to excess  $\text{Al}_2\text{O}_3$  (reflected by the presence of corundum) in the bulk compositions.

The calc-silicate assemblage with Cc-Qz-Gar-An-Wo appears to be invariant in the system  $\text{CaO--Al}_2\text{O}_3\text{--SiO}_2\text{--(CO}_2\text{)}$ , assuming diopside is due to an extra component of  $(\text{Fe,Mg})\text{O}$ . The textural and compositional characteristics of the phases are given below:

*Wollastonite* is in two generations. Acicular to prismatic crystals often intergrown with garnet are distinct from broad sub-idioblastic columnar crystals (0.5 to 1.00 mm) with ragged terminations. The second type also occurs as inclusions in garnet

with diopside, calcite, plagioclase, and quartz (fig. 2c). Compositionally, however, the two types of wollastonite are indistinguishable. They appear to be pure  $\text{CaSiO}_3$  with only 0.5 wt. % of  $\text{FeO} + \text{MgO} + \text{MnO}$ .

*Diopside* crystals are sub-equidimensional and occur in the groundmass as well as inside garnet. This is a salite with negligible Na and Al and with an  $\text{Mg}/(\text{Mg} + \text{Fe})$  ratio = 0.65.

*Garnet* occurs as porphyroblasts (3 to 4 mm diameter) containing inclusions of almost all the other main phases. It is mainly grossularite (90–4 vol. %) with 2–4 % andradite and 3.5 % almandine-pyrope.

Calcite, quartz, and plagioclase ( $\text{An}_{90}$ ) are minor constituents, which together do not make up more than 10 wt. % of the rock.

Based on field occurrence, petrography (see below), and modal mineralogy the garnet amphibolite was probably of metasedimentary origin:

*Hornblende* (15–20 %) occurs in the groundmass and as inclusions in garnet. The amphibole is uniform in composition, with  $\text{Al}^{\text{vi}}$  and Al (total) ranging from 0.7 to 0.75 and 2.8 to 2.9 cations per 23 oxygens, respectively. The  $\text{Fe}^{3+}/\text{Fe}^{2+}$  in the hornblende is 14 % and thus close to the values of amphiboles synthesized in tholeiitic basalt on the WM and QMF buffer (Spear, 1981). The  $X_{\text{Mg}}$  ratio  $[\text{Mg}/(\text{Mg} + \text{Fe}^{2+})]$  in the hornblende within garnet is, however, lower than that in the groundmass hornblende (Table I).

*Garnet* crystals are of two generations: large sub-rounded porphyroblasts (fig. 2d) containing inclusions of hornblende, chlorite, and epidote are an early generation; the second generation are small and euhedral and embedded in the chlorite matrix. Both types have significant spessartine contents (12–14 mol. %) with only 6–8 mol. % pyrope. The almandine and grossularite molecules range between 60–65 % and 20–22 %, respectively. These garnets have been analysed in detail and their compositional characteristics are discussed separately in the light of ambient *PT* conditions and of the possible reactions which caused these variations.

*Chlorite* laths are oriented at a high angle to the garnet porphyroblasts. Analyses of chlorite are given in Table I. In all three analyses chlorite has  $\text{Fe}/(\text{Fe} + \text{Mg})$  lower than that in coexisting garnet suggesting that it crystallized in equilibrium with it (cf. Thompson, 1976).

Data on *plagioclase* in Table I indicate about 20 mol. % anorthite.

*Epidote* is mostly restricted within garnet and is identified as zoisite  $[\text{Ca}_{2.09}\text{Al}_{2.73}\text{Fe}_{0.27}^{3+}\text{Si}_3\text{O}_{12}(\text{OH})]$  with somewhat excess Si in the tetrahedral site.

Table 1. Representative analyses (by electron probe) and formulae of minerals from "enclaves" within Banded Gneissic Complex, N.W.India

| Rock Type                      | Fuch-ky-Co Schist RS-17 |                   |        |                  |                  | Ctd Ky-Mus Schist, M20/740 |        |                 |                 | Calc-silicate rock, M20/596 |        |                 |                 | Gar-chl-amphibolite, M20/635 |                  |                  |                  |                 |        |  |  |  |  |
|--------------------------------|-------------------------|-------------------|--------|------------------|------------------|----------------------------|--------|-----------------|-----------------|-----------------------------|--------|-----------------|-----------------|------------------------------|------------------|------------------|------------------|-----------------|--------|--|--|--|--|
| Mineral                        | Fuch <sup>1</sup>       | Fuch <sup>2</sup> | Ky     | Mus <sup>1</sup> | Mus <sup>2</sup> | Ctd                        | Ru     | Wo <sup>3</sup> | Wo <sup>4</sup> | Diop                        | Gar    | Hb <sup>5</sup> | Hb <sup>6</sup> | Gar (Rim)                    | Chl <sup>5</sup> | Chl <sup>6</sup> | Chl <sup>7</sup> | Ep <sup>5</sup> | Plag   |  |  |  |  |
| SiO <sub>2</sub>               | 45.51                   | 46.72             | 36.53  | 46.34            | 46.76            | 23.96                      | 0.63   | 50.89           | 50.71           | 51.73                       | 38.99  | 42.84           | 41.56           | 36.82                        | 25.81            | 24.16            | 24.04            | 37.64           | 62.89  |  |  |  |  |
| TiO <sub>2</sub>               | 0.14                    | 0.31              | 0.07   | 0.07             | 0.00             | 0.03                       | 98.73  | 0.01            | 0.07            | 0.00                        | 0.29   | 0.15            | 0.08            | 0.07                         | 0.06             | 0.00             | 0.11             | -               | 0.04   |  |  |  |  |
| Al <sub>2</sub> O <sub>3</sub> | 36.72                   | 37.19             | 62.19  | 37.75            | 37.69            | 40.74                      | 0.13   | 0.20            | 0.00            | 0.23                        | 21.37  | 16.34           | 16.55           | 21.18                        | 22.72            | 22.62            | 22.37            | 28.22           | 23.02  |  |  |  |  |
| Cr <sub>2</sub> O <sub>3</sub> | 0.96                    | 0.75              | 0.31   | 0.15             | 0.06             | 0.07                       | 0.03   | 0.00            | 0.10            | 0.12                        | 0.04   | 0.04            | -               | -                            | 0.07             | -                | 0.12             | -               | -      |  |  |  |  |
| FeO <sub>t</sub>               | 0.96                    | 1.09              | 0.51   | 0.17             | 0.31             | 27.56                      | 0.08   | 0.28            | 0.26            | 9.49                        | 2.11   | 18.11           | 17.14           | 26.74                        | 26.24            | 29.72            | 29.27            | 6.52            | 0.04   |  |  |  |  |
| MnO                            | 0.04                    | 0.06              | 0.00   | -                | 0.01             | 0.18                       | 0.02   | 0.35            | 0.40            | 1.05                        | 0.65   | 0.16            | 0.11            | 5.31                         | 0.59             | 0.14             | 0.39             | 0.45            | -      |  |  |  |  |
| MgO                            | 0.26                    | 0.19              | -      | -                | 0.17             | 1.46                       | 0.03   | 0.14            | 0.00            | 11.85                       | 0.23   | 7.82            | 8.18            | 1.55                         | 14.40            | 11.72            | 11.64            | 0.28            | 0.00   |  |  |  |  |
| CaO                            | 0.08                    | 0.00              | -      | -                | -                | 0.07                       | 0.04   | 47.63           | 48.17           | 24.79                       | 36.33  | 11.53           | 11.74           | 7.92                         | 0.05             | 0.02             | 0.11             | 23.87           | 4.39   |  |  |  |  |
| Na <sub>2</sub> O              | 0.98                    | 0.77              | -      | 0.88             | 0.89             | 0.07                       | 0.18   | 0.04            | 0.04            | 0.36                        | 0.17   | 1.23            | 1.30            | 0.07                         | 0.23             | 0.34             | 0.16             | 0.31            | 8.88   |  |  |  |  |
| K <sub>2</sub> O               | 9.82                    | 10.24             | -      | 9.97             | 10.17            | 0.06                       | 0.00   | 0.05            | 0.00            | 0.00                        | -      | 0.48            | 0.45            | -                            | 0.02             | -                | 0.03             | -               | 0.07   |  |  |  |  |
|                                | 95.47                   | 97.32             | 99.61  | 95.33            | 96.06            | 94.20                      | 99.87  | 99.59           | 99.75           | 99.62                       | 100.18 | 98.70           | 97.11           | 99.66                        | 90.19            | 88.72            | 88.24            | 97.29           | 99.33  |  |  |  |  |
| Formula based on               |                         |                   |        |                  |                  |                            |        |                 |                 |                             |        |                 |                 |                              |                  |                  |                  |                 |        |  |  |  |  |
|                                | 22(O)                   | 22(O)             | 5(O)   | 22(O)            | 22(O)            | 12(O)                      | 2(O)   | 18(O)           | 18(O)           | 6(O)                        | 24(O)  | 23(O)           | 23(O)           | 24(O)                        | 28(O)            | 28(O)            | 28(O)            | 13(O)           | 32(O)  |  |  |  |  |
| Si                             | 6.037                   | 6.079             | 1.002  | 6.102            | 6.117            | 1.974                      | 0.008  | 5.948           | 5.931           | 1.972                       | 5.922  | 6.289           | 6.185           | 5.934                        | 5.307            | 5.159            | 5.170            | 3.101           | 11.186 |  |  |  |  |
| Al <sup>iv</sup>               | 1.963                   | 1.921             | -      | 1.898            | 1.883            | -                          | -      | 0.027           | -               | -                           | 0.078  | 1.711           | 1.815           | 0.066                        | 2.693            | 2.841            | 2.830            | -               | 0.814  |  |  |  |  |
| Al <sup>vi</sup>               | 3.779                   | 3.784             | 1.979  | 3.861            | 3.921            | 3.958                      | 0.002  | -               | -               | 0.010                       | 3.747  | 1.117           | 1.089           | 3.957                        | 2.813            | 2.851            | 2.839            | 2.749           | 4.011  |  |  |  |  |
| Cr                             | 0.100                   | 0.078             | 0.007  | 0.016            | 0.007            | 0.005                      | 0.00   | -               | 0.009           | 0.004                       | 0.004  | 0.005           | -               | -                            | 0.011            | -                | 0.020            | -               | 0.005  |  |  |  |  |
| Ti                             | 0.014                   | 0.031             | 0.001  | 0.007            | -                | 0.001                      | 0.987  | 0.001           | 0.002           | 0.000                       | 0.033  | 0.017           | 0.009           | 0.066                        | 0.010            | -                | 0.018            | -               | -      |  |  |  |  |
| Fe <sup>3+</sup>               | -                       | -                 | 0.013* | -                | -                | -                          | 0.001* | -               | -               | -                           | -      | 0.322*          | 0.384*          | -                            | -                | -                | -                | 0.499*          | -      |  |  |  |  |
| Fe <sup>2+</sup>               | 0.106*                  | 0.118*            | -      | 0.019*           | 0.034*           | 1.898*                     | -      | 0.027*          | 0.024*          | 0.303*                      | 0.268* | 1.902           | 1.749           | 3.604*                       | 4.512*           | 5.307*           | 5.263*           | -               | 0.007* |  |  |  |  |
| Mn                             | 0.004                   | 0.006             | -      | -                | 0.002            | 0.013                      | 0.000  | 0.035           | 0.019           | 0.034                       | 0.084  | 0.020           | 0.014           | 0.725                        | 0.104            | 0.026            | 0.054            | 0.032           | -      |  |  |  |  |
| Mg                             | 0.052                   | 0.035             | -      | -                | 0.033            | 0.180                      | 0.001  | 0.024           | -               | 0.673                       | 0.052  | 1.711           | 1.814           | 0.372                        | 4.414            | 3.730            | 3.730            | 0.035           | -      |  |  |  |  |
| Ca                             | 0.012                   | -                 | -      | -                | -                | 0.007                      | 0.001  | 5.964           | 6.039           | 1.013                       | 5.911  | 1.814           | 1.872           | 1.367                        | 0.012            | 0.003            | 0.026            | 2.114           | 0.835  |  |  |  |  |
| Na                             | 0.253                   | 0.195             | -      | 0.224            | 0.227            | 0.012                      | 0.005  | 0.009           | 0.009           | 0.027                       | 0.051  | 0.350           | 0.375           | 0.022                        | 0.090            | 0.140            | 0.068            | 0.050           | 3.066  |  |  |  |  |
| K                              | 1.662                   | 1.700             | -      | 1.675            | 1.699            | 0.007                      | -      | 0.008           | -               | 0.000                       | -      | 0.090           | 0.085           | -                            | -                | -                | -                | -               | 0.018  |  |  |  |  |

1. Oriented, large lath; 2. Small, random flake; 3. Long, slender prism interwoven with garnet; 4. Broad prism in groundmass; 5. Included within garnet; 6. Constituent phase of groundmass; 7. From a chlorite zone of about 0.5mm between the two garnet porphyroblasts.

\* Recalculated on the basis of stoichiometry.

\* Total iron as Fe<sup>2+</sup>

Amongst the accessories, ilmenite is homogeneous in composition and does not show exsolution of magnetite. From its structural formula  $[(\text{Fe}_{1.98}\text{Mn}_{0.02}\text{Mg}_{0.01})\text{Ti}_{2.08}\text{O}_3]$  it appears to be pure, possibly with very little ferric iron. Sphene is present but was not analysed.

### Garnet zoning

Field and petrographic evidence and the estimates of *PT* conditions (see following section) suggest that the rocks had recrystallized in the upper amphibolite facies and later in the greenschist facies. But did these two recrystallizations take place within a single orogenic cycle or during two separate events? In order to answer this question we have studied in detail an amphibolite which has well-developed chlorite in which small, euhedral garnets contrast with large subidioblastic garnets that are rimmed by chlorite (fig. 2d). Clearly chemical changes have taken place in the garnets, in which the *X*-site elements have responded

sensitively to the changing *PT* conditions and to the compositional changes of other phases involved in the garnet growth.

Garnets from sample M20/635 were scanned from one margin to the other via the core. The euhedral garnets (fig. 3a) show a bell-shaped Mn-profile similar to that described by Hollister (1966), and Fe and Mg show a decrease from a margin to core. These characteristics suggest there was a preferential incorporation of ions from the surrounding chlorite matrix which acted as a homogeneous reservoir for Mn (cf. Hollister, 1966; Atherton, 1968).

The larger crystals of garnet, however, do not show a simple profile for the *X*-site cations. Along the traverse line there are humps and depressions (fig. 3b) which probably indicate retrogression and recrystallization of the garnet; garnet growth has been reported during cooling (Atherton, 1968). We suggest there was a non-continuous growth of garnet due to polymetamorphism (Edmunds and Atherton, 1971) for which there is independent

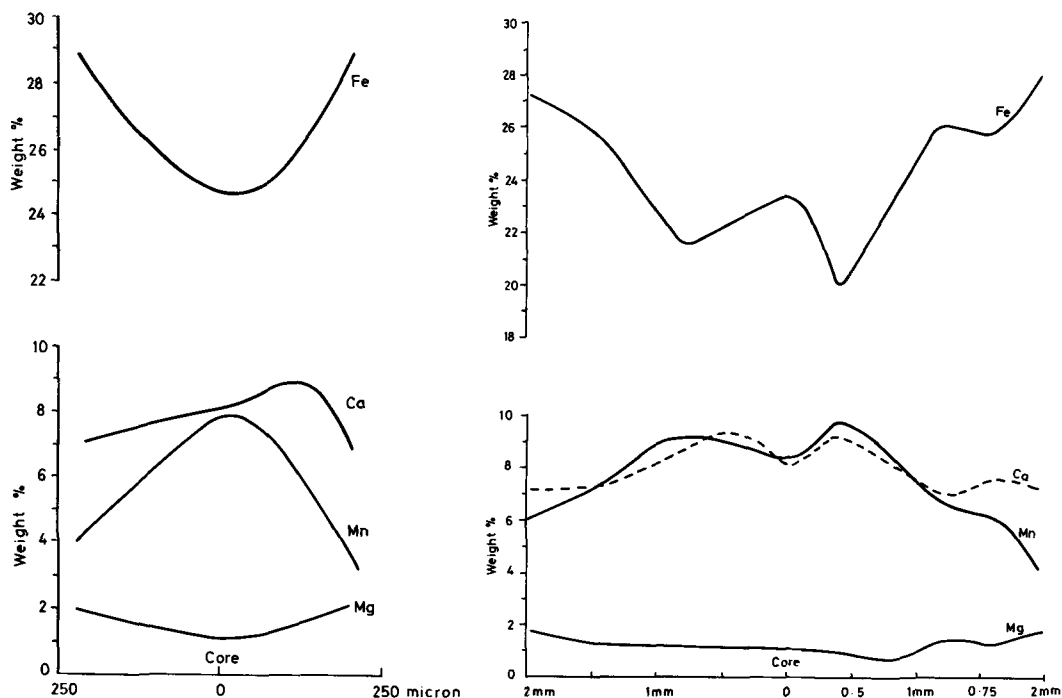


FIG. 3. Rim-to-rim compositional profiles for eightfold site elements (wt. %) plotted against distance in (a, left) garnet as a small euhedral crystal embedded in chlorite matrix and (b, right) garnet porphyroblast with subhedral outline in amphibolite (sample M20/635).

evidence in the rocks of the Banded Gneissic Complex in Rajasthan (cf Heron, 1953; Sharma and Narayan, 1975).

In order to study the detailed thermal history of the rocks a large garnet of 4 mm diameter was thoroughly analysed and chemical maps constructed (fig. 4). Its core is characterized by high MnO (about 10 wt. %) with an irregular zonation as obtained from the radial profiles (fig. 4a). The FeO distribution is complementary to that of MnO and reverse zoning is noticeable (fig. 4b). The chemical zoning of CaO is like that of MnO decreasing at the rim. However, reverse zoning by Ca and Mg is conspicuous in the intermediate region (figs 4c and d). The Mg content of the garnet is notably low compared with that of most amphibolites in regional metamorphic terrains.

Absence of hematite and the presence of ilmenite in the amphibolite negate the possibility that  $f_{O_2}$  was responsible for the observed garnet composition variations (Müller and Schneider, 1971).

Reverse zoning in Mn may be due to (a) decrease in growth rate (Edmunds and Atherton, 1971); (b) resorption of garnet and back-diffusion of Mn during prograde conditions (Evans and Guidotti, 1966; Birk 1973; Béthune *et al.*, 1975); (c) retro-

gression during which garnet releases Mg and Fe for chlorite but Mn, preferring the garnet structure, diffuses back into the garnet rim (Amit, 1976); (d) retrogression due to volatile introduction, wherein once homogeneous garnet was absorbed with a decrease in Mg/Fe ratio and increase in Mn-content at the garnet rim (Grant and Weiblen, 1971).

A literature survey reveals that garnet zoning may be explained as a function of bulk composition, reaction divariancy,  $P$  and  $T$ , fluids, modal amount of garnet, diffusion rates, and partition coefficients between garnet and related phases (see Tracy *et al.*, 1976). In order to constrain the choice of such possible mechanisms we have obtained data on morphology and on its compositional variations in relation to the surrounding and included minerals:

1. The core or the central region of the garnets is free of inclusions. This excludes rapid growth rate and syn-kinematic growth of garnet in its earlier stages.
2. The garnets contain inclusions of epidote, chlorite, and hornblende in their border zones (the latter two minerals are the main constituents of the

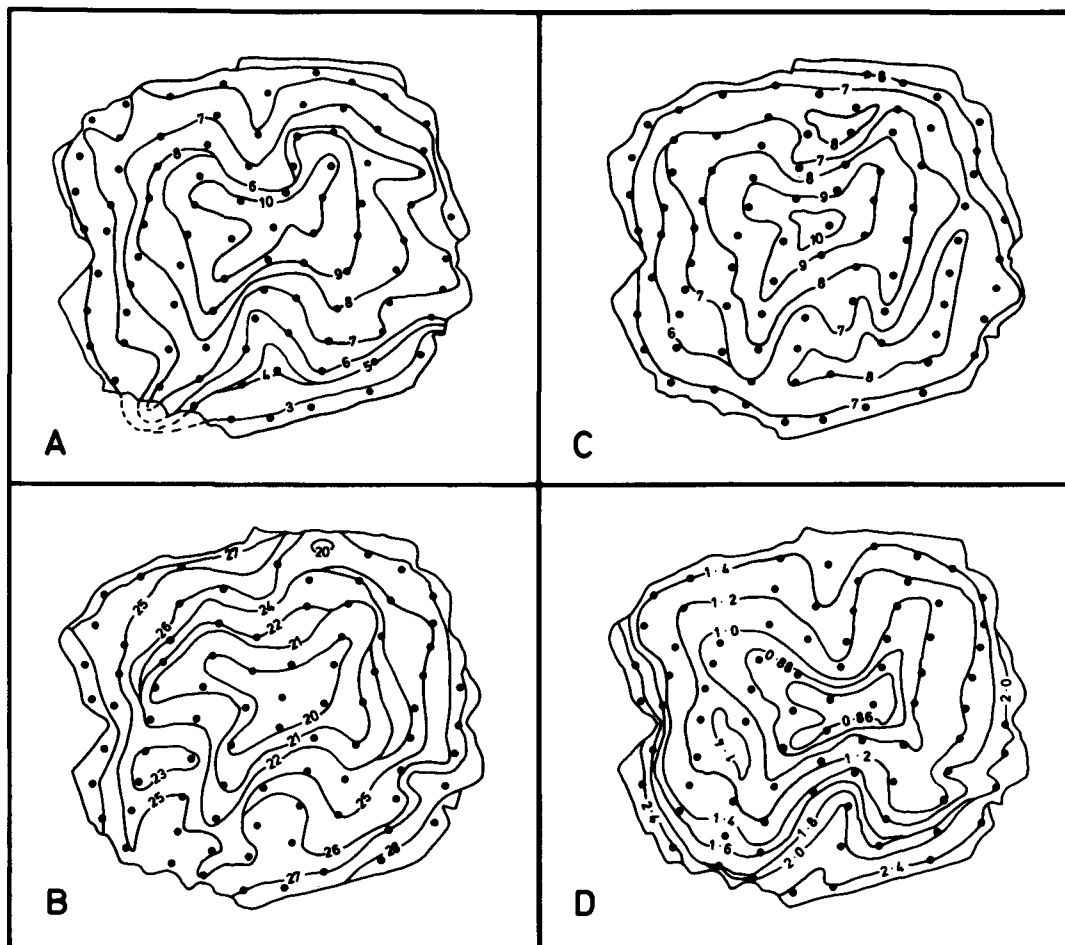


FIG. 4. Contours of Mn, Fe, Ca, and Mg (wt. %) in a garnet from amphibolite (sample No. M20/635). Dots indicate microprobe analyses spots.

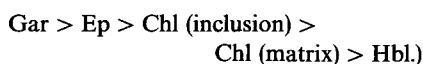
groundmass). This implies that progressive reaction has eliminated these phases from the garnet core; but later on conditions changed to make these minerals, mainly hornblende and chlorite, a stable paragenesis with garnet, i.e. progressive metamorphism was followed by retrogression.

3. Two garnet porphyroblasts (4 mm diameter), each with 3 wt. % MnO at the rim, are separated by a 500  $\mu$ m thin zone of chlorite containing 0.5 wt. % MnO. This feature excludes intergrain diffusion but requires a single precursor material—chlorite in the present case. In other words, the compositional variations in the garnets cannot be due to a garnet-matrix partitioning.

4. The core of the scanned garnet porphyroblast contains 3% by weight more MnO than the core of a small euhedral garnet (about 7 wt. % MnO). Both

garnets are surrounded by chlorite and have identical  $X_{Mn}$  values. The homogeneous nature of the groundmass and simple mineralogy of the rock exclude different rates of growth and hence the compositional changes in the garnet cannot be explained by the modal amount of garnet in the rocks, i.e. host rock composition and partition coefficient parameters can be ruled out.

5. Chlorite within garnet has a higher Mn content than chlorite in the matrix (see Table I). The  $X_{Mn}$  of the relevant phases in the rock has the following order:



The large spessartine content of the garnet core, the higher  $X_{Mn}$  in the included chlorite than in the

matrix chlorite, and the almost exclusive occurrence of epidote within garnet all suggest that most Mn was taken from chlorite and epidote which were the main reactants for the initial garnet-growth reaction.

Accordingly the chlorite was probably a mangiferous chamositic septechlorite and the epidote was a piedmontite type. If we assume that crystallization continued until the rock had reached the conditions of the upper amphibolite facies (indicated by the pelitic and psammitic gneisses and calc-silicate mineralogy) the chlorite must have disappeared (Hsu, 1968) long before the granite melting curve and the upper stability curve of muscovite + quartz were reached. Although the presence of quartz does not affect the thermal stability limit of Fe-chlorite (Hsu, 1968), Fawcett and Yoder (1966) claimed that excess  $\text{SiO}_2$  reduces the upper stability limit of Mg-chlorite about  $150^\circ\text{C}$  at 2 kbar fluid pressure and that the maximum temperature of the assemblage Mg-Chl-Qz is highly dependent on the Mg/Al ratio of the chlorite. Epidote as an alternative Mn-supplier disappears in basaltic compositions in runs over  $600^\circ\text{C}$  at a pressure greater than 2 kbar (Liou *et al.*, 1974).

Fig. 5 gives a schematic *PT* diagram for equilibrium involving garnet-chlorite-hornblende-plagioclase-epidote-quartz whose actual composition is plotted in an ACF diagram and projected from plagioclase on an extended C-F line (fig. 5a). Referring the model of recrystallization and retrogression to the *P-T* grid (see fig. 5 inset) the (Plag) reaction at the initial stage would gradually enrich the growing garnet in grossularite-spessartine components. This is because  $X_{\text{Fe}}$  in hornblende would rise with increasing temperature and the plagioclase instability at the initial conditions would enrich the coexisting garnet in Ca, as is reflected in the analysed garnet (see fig. 4a and c). Once plagioclase is formed as in the (Hb) reaction the grossularite component decreases towards the periphery. With a further rise in temperature the (Chl) reaction required that the Mn content of the garnet should decrease rather than increase, because the garnet would now be decomposed. Since the (Chl) reaction (divariant) would occur over a wide range of temperatures, garnet would not be eliminated and the reaction would be replaced by a discontinuous reaction:



This would stop when one of the reactant phases is consumed and the conditions are appropriate for chlorite to disappear, as stated earlier. The presence of hornblende, chlorite, and epidote within garnet suggests that the discontinuous reaction was not surpassed. It is most likely that the univariant

reaction in Fe-Mn-Ca become divariant in the Fe-Mn-Ca-Mg system.

The formation of plagioclase in the above-stated continuous or discontinuous reaction would decrease the grossularite content of the garnet, because the anorthite component in plagioclase increases with increasing temperature (Leake, 1965; Wenk and Keller, 1969). However, the presence of an 8% contour of CaO between the 7% CaO contours and the periphery (fig. 4c) indicates that there was once a sudden increase in temperature before the final cooling. This implies that garnet growth and resorption are reflected in the observed truncated compositional contours particularly for Fe and Mn (see fig. 4a and b). A late-stage retrogression is documented by the extensive development of chlorite, by the observed irregular outline of garnet porphyroblasts, by the truncated compositional contours, and by the fairly steep gradient in element concentration near the rim. The garnet dissolution and the development of chlorite would be facilitated at high pressure at the (Hb) reaction, and at a low pressure by the (Ep) reaction, as can be appreciated by the schematic grid (fig. 5b).

In this model it is impossible to decide at what stage the smaller euhedral garnets were formed, since they are embedded in the matrix formed of chlorite. However, it can be argued that *PT* changes were still appropriate for garnet growth on the garnet-chlorite equilibria (Hsu, 1968), and the bell-shaped Mn profile (fig. 4a) may be evidence of the garnet growth from the homogeneous matrix of chlorite (cf. Atherton, 1968). This would imply that the rim of large garnets resulted from a prograde reaction (because Mn depletion during a retrograde stage would require garnet growth, but this growth is a prograde phenomenon, cf. Novak and Holdaway, 1981) and hence an alternative model is required to explain the observed zoning in garnets.

The alternative model presented below involves two recrystallization periods, separated in space and time. We prefer it to the first model since the compositional characteristics of large garnets cannot be satisfactorily explained as a function of a single prograde metamorphism and the rocks of the BGC are polymetamorphic *sensu stricto* (Heron, 1953; Sharma, 1977; Sharma and MacRae, 1981). According to this model the garnets in the amphibolite are of two generations: early regional large crystals ( $\text{Gt}_1$ ); and late regional small, euhedral crystals ( $\text{Gt}_2$ ) whose compositional characteristics seem to be a function of a single prograde metamorphic process. In other words, an early garnet of M1 period was resorbed during the initial stages of M2 metamorphism, after which its growth continued again until close to the peak of meta-

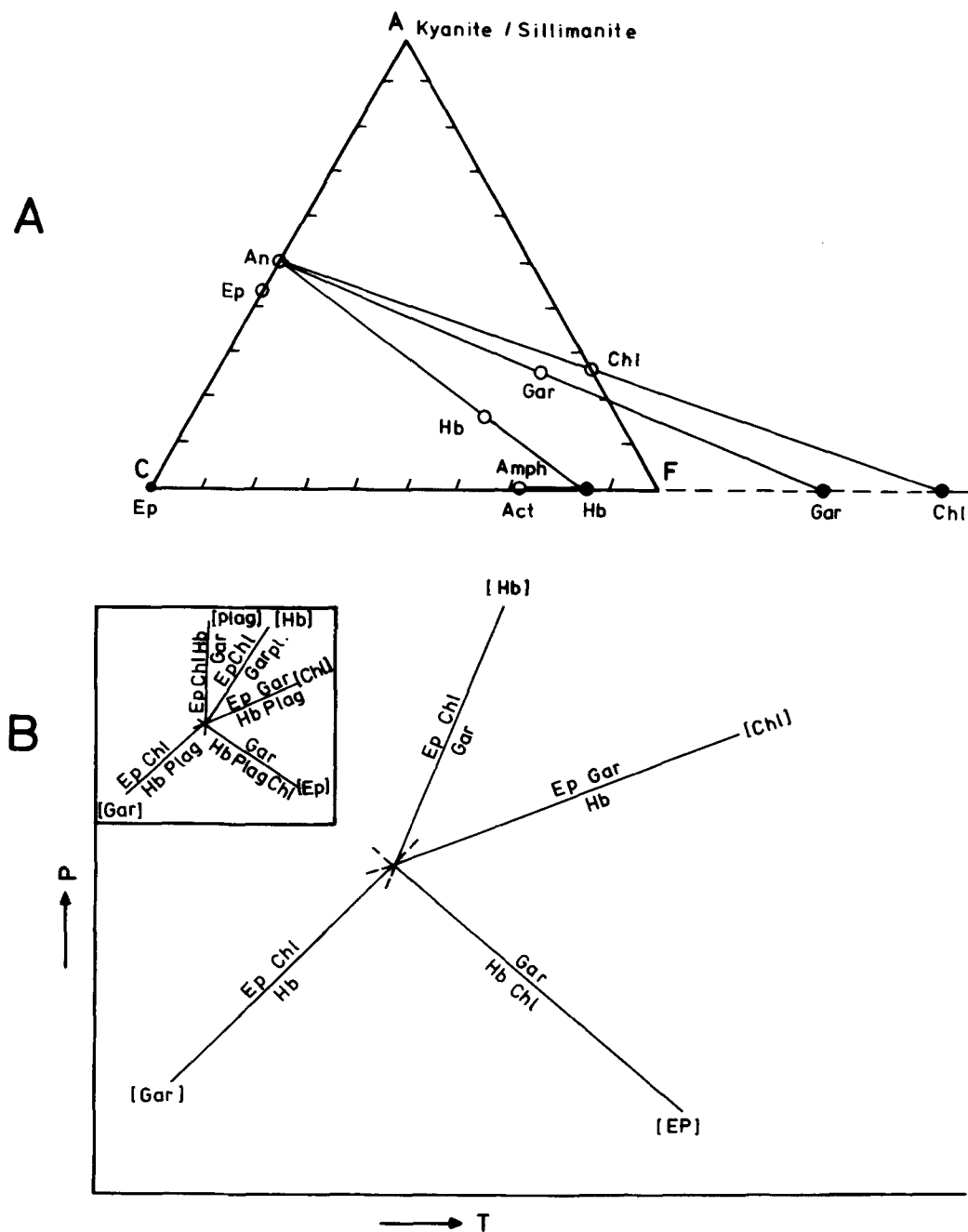


Fig. 5. (a) The analysed phases of the basic assemblages are shown in an ACF diagram and also after projection on C-F line (extended) from anorthite. Abbreviations: Act = Actinolite; An = anorthite; Chl = chlorite; Ep = epidote; Gar = garnet; Hb = hornblende. (b) A schematic petrogenetic grid relating the phases chlorite-garnet-hornblende-epidote in association with plagioclase and quartz and with excess  $H_2O$ . The standard state data for the reactions are from Wells (1979, Table 1) and Robie *et al.* (1979). Inset shows appearance of plagioclase in the different divariant reactions and aids in predicting  $X_{ca}$  variation mainly in the coexisting garnet.

morphism. Thereafter a stage of retrogression occurred which formed chlorite. This model conforms with the polymetamorphic character of the gneissic complex and can be appreciated from the  $P$ - $T$  grid (fig. 5b) in which we notice the following sequence of events. (a) An early garnet is resorbed at the (Gar) reaction. (b) With increasing pressure and temperature the garnet rim would recrystallize and at the same time euhedral garnet would develop at the (Ep) reaction. The recrystallization or garnet growth would continue up to the (Chl) reaction. (c) During the cooling stage chlorite formation would take place such as at the (Hb) reaction.

This garnet relationship is analogous to that from the Fanad aureole in Donegal (Edmunds and Atherton, 1971). The second recrystallization is not associated with an intrusive body as in Fanad, but is here caused by regional metamorphism (M2) to which most of the Gneissic Complex was subjected (cf. Heron, 1953; Sharma, 1982). During the initial stage of superimposed metamorphism (M2 = Aravalli-Delhi orogeny; 1650 Ma. Sharma, 1982; Crawford, 1970) the early garnet  $Gt_1$  underwent resorption, which gave rise to intergranular diffusion. In order to stabilize the garnet, Mn would be concentrated in the core and be greater than that released with Fe and Mg to form chlorite. This explains the extra concentration of Mn in the garnet cores and does not demand an unusual composition of sediments or minerals involved in garnet formation. Complete retrogression did not occur unlike the Fanad garnets, Donegal (Edmunds and Atherton, 1971), because the second recrystallization was also at regional medium pressures above 6 kbar (Sharma and MacRae, 1981). Also a more Mg-rich, in contrast to Fe-rich, composition would have caused complete resorption. (Thompson *et al.*, 1977).

The concentration of MnO, FeO, and CaO at the edges of  $Gt_1$  and in the rim of  $Gt_2$  is almost the same. Moreover, the core composition of  $Gt_2$  corresponds closely to the composition at the intermediate region of  $Gt_1$ . These relations suggest that a homogeneous reservoir (chlorite) was available to the resorbed garnet and the newly crystallized euhedral garnet during the second metamorphism. The reactions apparently continued until all MnO from the groundmass chlorite had been consumed. The epidote-chlorite-hornblende inclusions near the periphery of  $Gt_1$  testify to the presence of prograde metamorphic conditions under which the garnet rim has grown. This non-continuous growth is revealed by the reverse zoning of Ca, Fe, and Mn in the late intermediate regions of  $Gt_1$ . The interior of  $Gt_1$  appears to have been out of equilibrium with the matrix during the

successive metamorphic event in a similar manner to the garnets from a garnet-mica schist in Inverness-shire (McAteer, 1976; see also Atherton and Edmunds, 1966).

### Metamorphic conditions

In view of the widespread but partial retrogression in pelitic/basic assemblages and associated gneisses, we are able to calculate temperature conditions of the regional metamorphism. The staurolite relict within chloritoid of the chloritoid-kyanite-quartz-muscovite assemblage indicates temperatures in the range 550–650 °C when the reactions:  $Fe-Ctd + Ky = Sta + Qz + Fluid$  (Richardson, 1968) and  $Fe-Ctd + Qz = Sta + Alm + Fluid$  (Ganguly, 1969; Hoschek, 1969) are considered (fig. 6). Because of the Fe-rich character of the chloritoid, staurolite could not coexist with garnet, even if the latter was present in the chloritoid-kyanite-bearing assemblage (cf. Thompson *et al.*, 1977, p. 1161). Because the phases have appreciable Mg contents, the temperature must have been higher than for the Fe end-member reactions of these authors. Similar temperatures are also obtained when garnet stability in the reaction  $Fe-Chl + Qz + Mt = Alm + Fluid$  (Hsu, 1968) is considered.

However, the equilibrium temperature of these reactions varies with the values of  $f_{O_2}$  during metamorphism. Absence of hematite and more frequent occurrence of ilmenite in chloritoid-bearing assemblages and in the metabasites favour low  $f_{O_2}$ , not above that within the magnetite stability field (cf. Graham, 1974). Temperatures higher than 650 °C at regional pressures could have existed, because sillimanite relicts occur within muscovite of the widespread quartz-plagioclase-K-feldspar-mica gneiss of the region. However, the equilibrium temperature of the reaction ( $Mus + Qz = K-felds + Al_2SiO_5 + H_2O$ ; Chatterjee and Johannes, 1974) is displaced to lower temperatures if  $P_{H_2O}$  is less than  $P$  total. The  $X_{H_2O}$  value would not have been below 0.5, especially when the calc-silicate bands have insignificant volume in the area.

Prevalence of such high temperatures (650–700 °C, depending on  $x_{H_2O}$  values of the fluid phase) is also supported by the calcite-quartz-wollastonite-grossularite-clinopyroxene assemblage, although the pertinent reactions,  $Cc + Qz = Wo + CO_2$  and  $Gross + Qz = An + Wo$ , are much influenced by the  $CO_2$  content in the gas phase and prevailing load pressure (cf. Greenwood, 1967; Gordon and Greenwood, 1971). If we displace the wollastonite-forming reactions so they intersect the upper stability curve of  $Stau + Qz$  (not coexisting in pelitic schist of the BGC), the required curve would

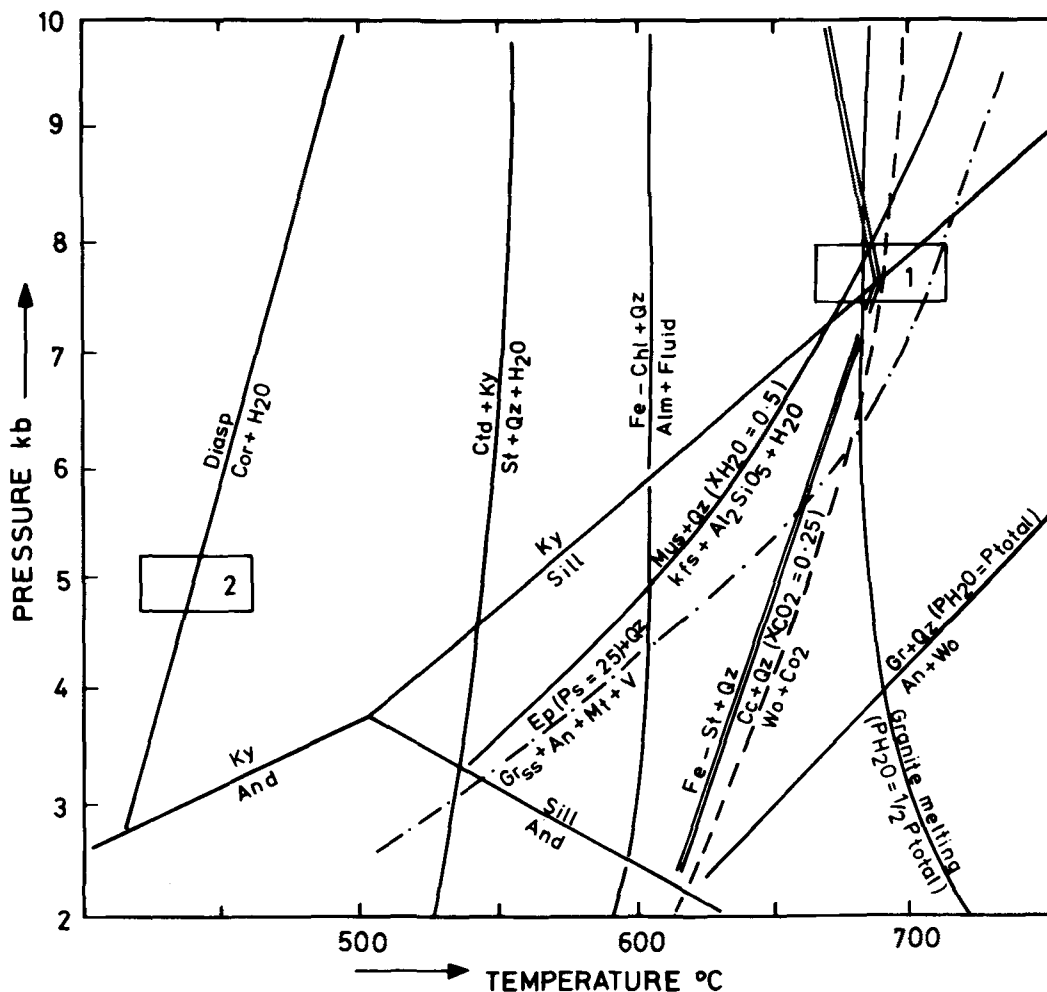


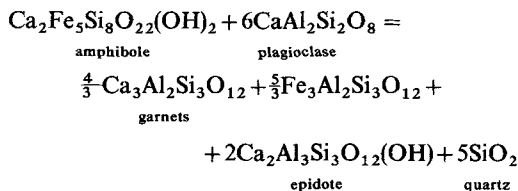
FIG. 6.  $P$ - $T$  diagram showing experimentally determined and thermodynamically calculated mineral equilibria pertinent to the present study. Triple point of  $\text{Al}_2\text{SiO}_5$  and the univariant equilibria of the aluminium silicate phases are after Holdaway (1971). For remaining curves see text. The two rectangular areas denote possible maximum  $PT$  conditions attained during last regional metamorphism (= Aravalli orogeny; 1650-950 Ma) as documented in the mineralogy of the basement gneiss complex and its metasedimentary enclaves. For discussion see text.

have an  $X_{\text{CO}_2}$  of 0.25. This is plausible, since calc-silicates only form small lenses in the gneisses. In the basic compositions epidote occurs mainly as inclusions within garnet and is rare in the matrix of the hornblende-garnet-plagioclase-chlorite  $\pm$  quartz assemblage. This may be taken to suggest that temperatures were in excess of 650 °C when the epidote  $\pm$  quartz stability (Liou, 1973) is referred to (see fig. 6). It is interesting to note that the Mus + Qz reaction at  $X_{\text{H}_2\text{O}} = 0.5$  (after Kerrick, 1972) intersects the upper thermal stability of Stau + Qz (curve at higher pressure is thermo-

dynamically calculated by Yardley, 1981) at about 700 °C, meeting the granite-melting curve of  $P_{\text{H}_2\text{O}} = 0.5P$  total (see fig. 6). This is relevant, because migmatites are observed in the field, although their gneissic component contains muscovite which is probably due to subsequent retrogression, as will be discussed later.

Pressure at such a temperature appears to be close to, or slightly above, the ky = sill univariant curve (Holdaway, 1971), since kyanite is persistently found in the aluminous-rich compositions, whereas sillimanite is only a relict phase in the enclosing

gneisses. The phases in the basic compositions seem to be related (despite abundance of chlorite) by the reaction:



for which equation (4) of Wells (1979) can be used for a pressure calculation. The required geobarometric expression is:

$$P \text{ (kbar)} = \frac{(28.46 + \ln K) T - 9494}{1.72128} + 1$$

where  $K = \frac{\text{epidote}}{\text{garnet}}$

$$\frac{(a_{\text{Ca}_2\text{Al}_2\text{Si}_2\text{O}_8(\text{OH})})^2 (a^4_{\text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12}} \cdot a^5_{\text{Fe}_3\text{Al}_2\text{Si}_3\text{O}_{12}})}{(a_{\text{Ca}_2\text{Fe}_5\text{Si}_8\text{O}_{22}(\text{OH})_2}) \cdot (a_{\text{CaAl}_2\text{Si}_2\text{O}_8})^6}$$

amphibole                      plagioclase

The pressure uncertainty is  $\pm 1.8$  kbar.

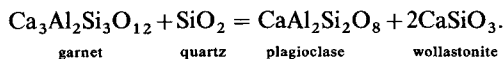
Activities for components participating in the stated equilibria have been calculated from mineral compositions as suggested by Wells (1979, pp. 665–6).

We calculate the equilibrium pressure at 700 °C to be 9.8 kbar, when ideal mixing is taken for all phases including garnet and plagioclase. However, the activity consideration in garnet (Ganguly and Kennedy, 1974) and plagioclase (Orville, 1972) lowers the pressure value to 7.2 kbar. The calibration based on more recent data on activity coefficients in garnet and plagioclase (Newton and Haselton, 1981) yields a pressure of 7 kbar, which is almost identical to that obtained from  $a$  to  $X$  data for garnet by Ganguly and Kennedy (1974) and from activity coefficients for plagioclase by Orville (1972). This coincidence may be understood from the fact that the activity–composition relations of Ca–Mg–Fe garnets deduced from Ganguly and Kennedy and the mixing properties of the ternary garnets derived calorimetrically by Newton and Haselton (1981) yield almost identical values at a recrystallization temperature of 700 °C for the investigated rocks. Moreover, the calculation of the plagioclase activity coefficient using the  $W$  values of the latter workers and after Orville gives similar  $a_{\text{An}}^{\text{Plag}}$  values of 1.32 cal mol.<sup>–1</sup> and 1.28 cal mol.<sup>–1</sup>, respectively.

The intersection of the granite melting curve (Piwinski and Wyllie, 1970) at  $P_{\text{H}_2\text{O}} = 0.5$  total and the Mus-involving reactions at  $X_{\text{H}_2\text{O}} = 0.5$  with the Fe–Staur + Qz stability and K–Sill curve at about 8 kbar (see fig. 6) corroborates the pressure estimates from the Wells (1979) equation. Thus

8 kbar seems to be a realistic pressure for the assemblages concerned.

We now relate the calc-silicate assemblage in sample M20/596 to the experimentally reversed reaction (Newton, 1966; Boettcher, 1970; Wisdom and Boettcher, 1976):



We made a pressure calculation from the relationship:

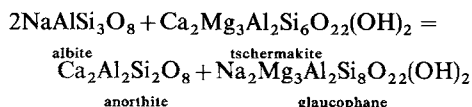
$$P_2 = P_1 - (RT \ln K) / \Delta V_s,$$

where  $K = \frac{a_{\text{anorthite (plag)}}}{a_{\text{grossularite (garnet)}}$

and  $P_2$  and  $P_1$  are the equilibrium pressures (in bars) of the reaction for impure and pure phases, respectively, at temperature  $T$  (°K).  $P_1$  was calculated from the relation  $P_{(\text{bars})} = 25.5 (T^\circ\text{C}) - 13300$ , derived by Droop and Treloar (1981) from the experimental brackets on the above reaction involving pure phases. Taking  $\Delta V_s$  from Robie *et al.* (1979), and assuming ideal mixing, the pressure estimate at 700 °C is close to 4 kbar. We propose that this represents the pressure during cooling because widespread retrogression suggests that the  $PT$  conditions of prograde metamorphism were appreciably lowered. In this event many or more of the reaction equilibria were reversed.

The maximum temperature during the cooling event must have been below the upper stability of Ctd + Ky, because this pair is stable in aluminous-rich compositions. The fuchsite–corundum–kyanite rock suffered no mineralogical changes during prograde or retrograde metamorphism, because of its simple bulk composition. However, the occurrence of a yellowish to dirty-white substance (? diaspore) around corundum near a quartz cluster in the fuchsite–kyanite rock may be taken to suggest that there was a low-temperature hydration close to the reaction Diaspore = Corundum + H<sub>2</sub>O (Haas, 1972) after cessation of tectonic activity. Pyrophyllite has not been identified, however. The absence of pyrophyllite can be attributed either to limited access of water and reaction rate phenomena (Fyfe *et al.*, 1978), or more likely to the fact that retrogressive metamorphism occurred at elevated pressures. In the phase diagram of Day (1976) there is an upper pressure stability limit for pyrophyllite which decomposes to diaspore + quartz in excess of 5 kbar. That the pressure during retrogression was fairly high is reflected in the absence of actinolite in the hornblende–plagioclase–garnet–chlorite assemblage. According to Misch and Rice (1975) and Oba (1980) actinolite–hornblende show complete miscibility at high pressures close to 5 kbar.

Quenching temperatures can be obtained from  $\text{NaSi} \rightleftharpoons \text{CaAl}$  exchange between hornblende and plagioclase in basic rocks. The phases can be related by the equation (Spear, 1980)



for which the equilibrium relation (Spear, 1980) is:

$$4700 - 44.27 T (^{\circ}\text{K}) - 0.22 P + 2RT \ln K = 0,$$

where 
$$K = \frac{X_{\text{An}} \cdot X_{\text{Na(M4)}}}{X_{\text{Ab}} \cdot X_{\text{Ca(M4)}}}$$

This yields temperatures of 506°C and 446°C, respectively, for hornblende within garnet and in the matrix. Although the plagioclase composition in our assemblage is 5% lower than that at which the above equation is valid, the temperature estimates should not differ by more than 25°C. Note that amphibolite with plagioclase containing  $\text{An}_{20}$  and  $\text{An}_{25}$  occurs in the same zone (cf. Spear, 1981, Table I). Also the  $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Fe}^{2+})$  ratio in these hornblendes agrees with that reported in most hornblendes and therefore we consider that the  $\text{Fe}^{3+}$  estimates are reasonable.

We must emphasize that retrogression was not the last stage of metamorphic re-equilibration in these rocks. This is based on the observation that kyanite retrogression has not given rise to shimmer aggregates, and coarse muscovite flakes often show preferred orientation. This implies that retrogression took place in a dynamic environment rather than during post-tectonic uplift. Such post-tectonic recrystallization is indicated by white mica after kyanite without any preferred orientation.

From the above discussion we conclude the final quenching was close to the Diaspore = Corundum +  $\text{H}_2\text{O}$  reaction curve at pressures of 4–5 kbar. This means that about 10 km of rock overload were removed by erosion.

#### *Implications for regional geology*

The early Proterozoic Aravalli meta-sediments were deposited unconformably on the Archaean BGC which contains enclaves of meta-supracrustals. Deformation associated with the Aravalli Orogeny affected the cover-basement surfaces infolding synclinal keels of Aravalli meta-sediments into the basement gneisses (Heron, 1953). Following uplift and erosion we see today that the BGC gneisses contain two types and ages of meta-supracrustal rocks—small metre-size early enclaves of calc-silicate rocks and meta-basics (garnet–hornblende–chlorite schists) and kilometre-long keels of Aravalli

kyanite-bearing schists. Heron (1953) and Sharma and Narayan (1975) showed that the BGC was a polymetamorphic, multideformed tectonic unit. The latter workers also presented evidence that medium-pressure kyanite-type metamorphism (up to the staurolite–kyanite zone) had occurred in Pre-Aravalli time and was later superimposed by another regional metamorphism associated with the Aravalli Orogeny which involved both the Basement Complex and the Aravalli supracrustals. The basement rocks were subjected in Aravalli times to a higher load pressure than they were during Pre-Aravalli time ( $M_1$ ) (Sharma, 1977). From the zonal distribution of minerals and paragenetic relationships Sharma (1977) and Sharma and MacRae (1981) established that sillimanite-grade metamorphism represents the second metamorphic episode ( $M_2$  = Aravalli orogenic cycle; 1650–950 Ma) associated with the high pressure. This overprinting is probably the cause of the younger mineral ages in the BGC rocks (cf. Crawford, 1970, and references therein; Sharma, 1982). Whether the thermal peak extended over the whole terrain of the Rajasthan Precambrian or whether it was of local extent, cannot be ascertained until a  $PT$  grid is constructed over a larger area. However, it is almost certain that the sillimanite event was followed by retrogressive metamorphism of varying degrees, and this implies that during the late phase of  $M_2$  metamorphism the activity of water and cooling temperatures varied in the basement rocks and their enclaves.

Although the kyanite-bearing, Al-rich meta-sedimentary keels show structural concordance with the surrounding BGC rocks, high temperatures close to the muscovite–quartz stability (evidenced by sillimanite relics in muscovite in BGC schist/gneisses and by the presence of anatectic migmatite) are not documented in their mineralogy. The co-existence of kyanite–chloritoid–quartz–muscovite and relics of staurolite observed in one locality only indicate conditions at or below lower amphibolite facies. Consideration of the compositional characteristics of the kyanite–chloritoid and kyanite–fuchsite schists, which are rich in  $\text{Al}_2\text{O}_3$  and Fe and low in  $\text{SiO}_2$ , CaO, MgO, and alkalis, leads us to accept that the protolith was a bauxitic clay derived by weathering of pre-existing rocks (cf. Morrison, 1972) which is consistent with the idea that the kyanite-bearing metasediments belong to a younger stratigraphic group.

The calc-silicate enclaves within the rocks of the basement complex are consistent with the high-temperature conditions that once existed in the BGC schists and gneisses. The two generations of wollastonite may be explained as follows: in the earlier metamorphism ( $M_1$ ) wollastonite was formed

by the reaction calcite + quartz = wollastonite + CO<sub>2</sub> and in the second (higher temperature) by the reaction: grossularite + quartz = anorthite + wollastonite.

Although amphibole-chlorite-epidote-quartz is a common assemblage in greenschists, amphibolites, and blueschists (Laird, 1980), the metabasic enclaves reveal their polymetamorphic history in the garnets. The discontinuously zoned garnets indicate multiple periods of growth, and this is consistent with poly-metamorphism. The evidence of the garnet zoning together with the fact that Na and Al in the hornblende of the groundmass are higher than in the hornblende included within garnet of the amphibolite are in complete accord with the geological evidence that higher pressures prevailed during the superimposed regional metamorphism (cf. Laird and Albee, 1981; see also Sharma and MacRae, 1981).

The pressure estimates of about 8 kbar indicate 24 km of overload during the prograde stage of the M<sub>2</sub> metamorphism. At this pressure the temperature was 700°C (fig. 6) and thus about 200°C higher than the temperature during retrogressive metamorphism when the pressure had dropped to 4–5 kbar. These estimates also indicate that about 12 km of overload were removed by the time the rocks were finally quenched. When more data on the age and *PT* conditions during the Aravalli orogeny are available, more realistic values for the geothermal gradient and the rate of erosion and uplift will be possible.

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