

# Petrology of Biotite–Cordierite–Garnet Gneiss of the McCullough Range, Nevada I. Evidence for Proterozoic Low-Pressure Fluid-Absent Granulite-Grade Metamorphism in the Southern Cordillera

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## ABSTRACT

Proterozoic migmatitic paragneisses exposed in the McCullough Range, southern Nevada, consist of cordierite + almanditic garnet + biotite + sillimanite + plagioclase + K-feldspar + quartz + ilmenite + hercynite. This assemblage is indicative of a low-pressure facies series at hornblende–granulite grade. Textures record a single metamorphic event involving crystallization of cordierite at the expense of biotite and sillimanite.

Thermobarometry utilizing cation exchange between garnet, biotite, cordierite, hercynite, and plagioclase yields a preferred temperature range of 590–750°C and a pressure range of 3–4 kb. Equilibrium among biotite, sillimanite, quartz, garnet, and K-feldspar records  $a_{H_2O}$  between 0.03 and 0.26. The low  $a_{H_2O}$  together with low  $f_{O_2}$  ( $\leq$  QFM) and optical properties of cordierite indicate metamorphism under fluid-absent conditions. Preserved mineral compositions are not consistent with equilibrium with a melt phase. Earlier limited partial melting was apparently extensive enough to cause desiccation of the pelitic assemblage.

The relatively low pressures attending high-grade metamorphism of the McCullough Range paragneisses allies this terrane with biotite–cordierite–garnet granulites in other orogenic belts. Closure pressures and temperatures require a transient apparent thermal gradient of at least 50 °C/km during part of this Proterozoic event in the southern Cordillera.

## INTRODUCTION

Numerous studies have documented granulite facies metamorphism that occurred under conditions of relatively high lithostatic pressures indicative of mid-crustal to lower-crustal levels (6–10 kb) (e.g., Grew, 1981; Martignole & Nantel, 1982; Tracy & Dietsch, 1982; Abscher & McSween, 1985; Bohlen *et al.*, 1985; Ellis & Green, 1985). However, several relatively low-pressure granulite terranes ( $\leq$  5 kb) have also been identified (e.g., Ashworth & Chinner, 1978; Ibarguchi & Martinez, 1982; Jamieson, 1984; Schreurs & Westra, 1986). These low-pressure granulites typically include metapelites with the assemblage biotite + cordierite + garnet + sillimanite + K-feldspar (kinzigites, following the usage of Volborth, 1973).

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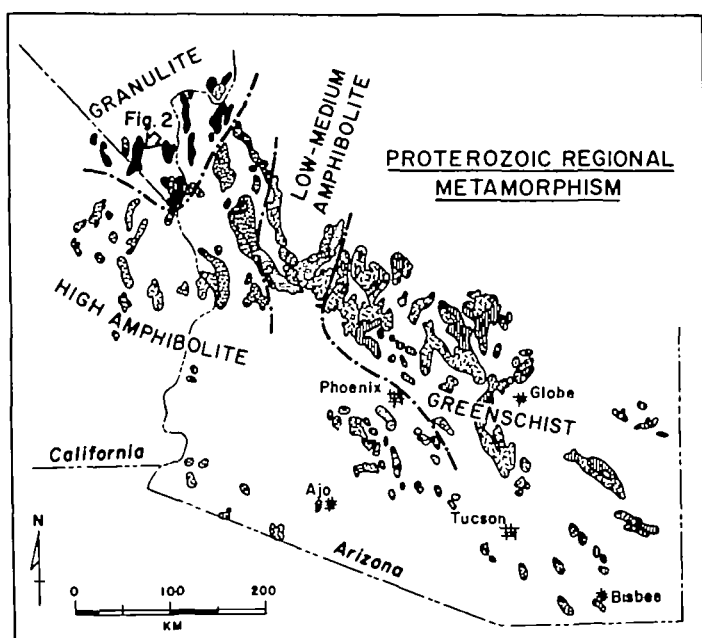


FIG. 1. Distribution of Early Proterozoic metamorphic grade in the lower Colorado River region and environs, after Anderson (1987). The heavy dot-dash lines indicate approximate boundaries of metamorphic grade but are not isograds in that they are polychronic. The symbols are: solid dark, granulite facies; heavy stipple, high amphibolite facies; vertical lines, greenschist facies; short lines, Proterozoic anorogenic granites.

Proterozoic kinzigite gneisses in the lower Colorado River region of southern Nevada, southeastern California, and adjacent Arizona were first described by Volborth (1962, 1973). Thomas *et al.* (1988) described the distribution and general petrological attributes of these high grade rocks (Fig. 1). This paper presents the first detailed analysis of the metamorphism based principally on the petrology of pelitic gneisses exposed in the McCullough Range of Southern Nevada. Geothermobarometric data are presented which indicate high-temperature low-pressure metamorphism of these Proterozoic rocks under fluid-absent conditions. The reaction history and  $P$ - $T$  path of the rocks is presented in a companion paper (Part II).

### GEOLOGIC SETTING AND CHRONOLOGY

The McCullough Range of southern Nevada comprises the northern extension of an Early Proterozoic terrane exposed in the northern New York and Ivanpah Mountains in adjacent northern California (Miller & Wooden, 1988). Recent mapping by Anderson *et al.* (1985) of a 85-km<sup>2</sup> area at a scale of 1:24 000 encompasses most of the crystalline terrane, and is summarized in Fig. 2. The crystalline rocks consist of four distinctive lithologies: (1) variably migmatitic kinzigite paragneiss; (2) lenses and boudins of amphibolite enclosed in paragneiss; (3) meta-ultramafic rocks interlayered with paragneiss and amphibolite; and (4) foliated to non-foliated granitic and dioritic plutons. With the exception of minor orthogneiss (see below), the intrusive rocks were not subjected to the high-grade metamorphic conditions evidenced by the surrounding country rock. However, the granitic plutons commonly exhibit a weakly developed foliation parallel with the north-northeast

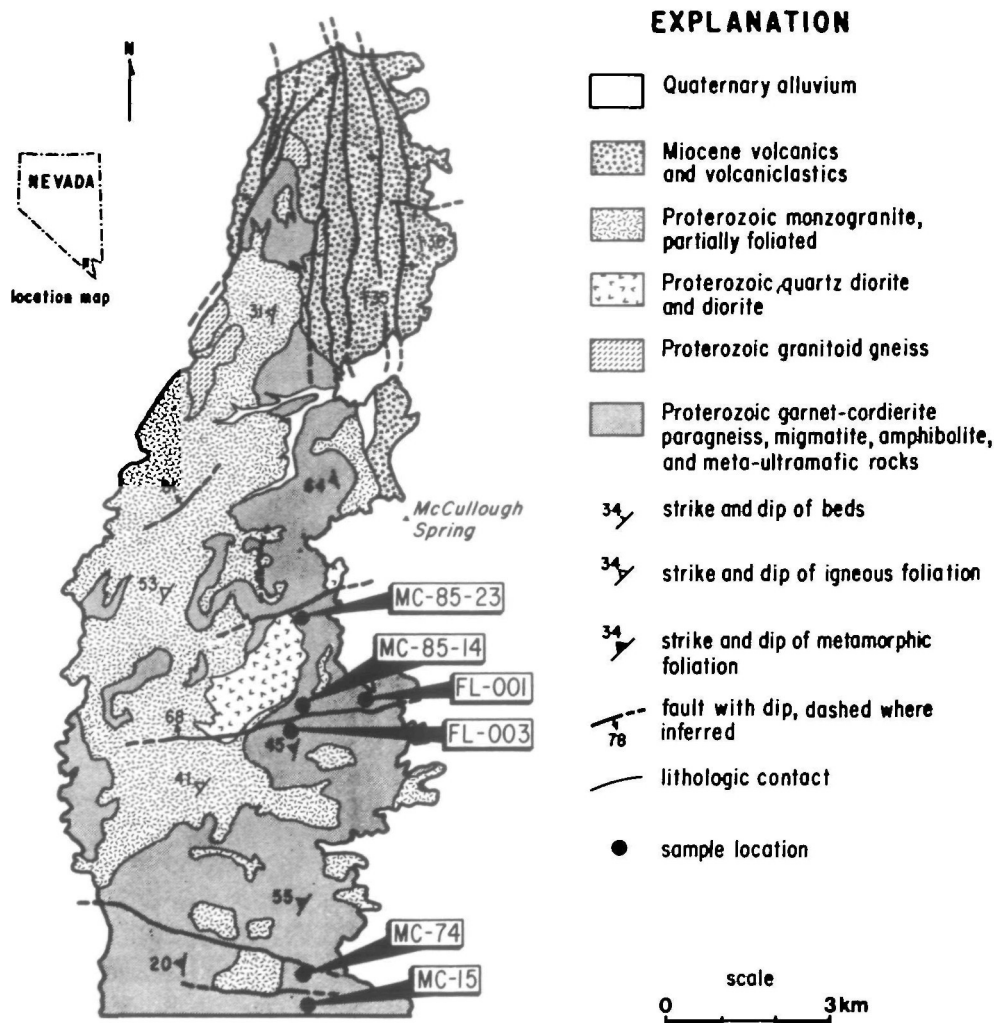


FIG. 2. Simplified geologic map of the southern McCullough Range, Nevada, showing the location of the samples discussed in text.

striking, west dipping  $S_1$  foliation of the paragneisses, and no contact thermal aureoles are present. Apparently, the granitoid plutons were emplaced during the waning stages of the  $S_1$ -forming event.

The crystalline complex of the McCullough Range is largely lithologically correlative with rocks exposed in the adjacent northern New York Mountains where Miller & Wooden (1988) have suggested an age of 1.71–1.70 Ga for high-grade metamorphism followed by post-kinematic granite intrusion at 1.69–1.66 Ga. Wooden (pers. comm.) has obtained a U–Pb zircon crystallization age of 1.71 Ga from an orthogneiss in the McCullough Range complex. The age of metamorphism in the range is thus constrained at 1.71–1.70 Ga.

Anorogenic granitic plutons of 1.4–1.5 Ga age, common in Proterozoic terranes of adjacent ranges and elsewhere in the North American continent (Anderson, 1983, 1987), are absent in the McCullough Range. K–Ar dates from biotite and hornblende mineral

separates from the McCullough complex range from  $1215 \pm 36$  Ma to  $790 \pm 24$  Ma (Anderson, 1986, unpublished data). The antiquity of these cooling ages is indicative of the absence of Phanerozoic thermo-tectonic overprinting, and a shallow crustal residence since Proterozoic time.

Post-granitoid mylonitic shear zones impart a foliation generally parallel to  $S_1$  in paragneisses and granitoids. Biotite from a sample of mylonitized monzogranite yields a K-Ar date of  $695 \pm 21$  Ma indicating shearing was also a Proterozoic event.

### GRADE OF METAMORPHISM

The assemblages garnet + sillimanite + cordierite + biotite + K-feldspar in the pelitic gneisses, orthopyroxene + actinolitic hornblende in meta-ultramafic rocks, and clinopyroxene + orthopyroxene + cummingtonite in amphibolites, suggest metamorphic conditions corresponding to the hornblende-granulite subfacies (deWaard, 1966; Green & Ringwood, 1967; Abscher & McSween, 1985). Coexistence of cordierite + garnet + biotite is indicative of a low-pressure facies series at this grade (deWaard, 1966; Green & Ringwood, 1967; Turner, 1968). The occurrence of cummingtonite instead of garnet in the amphibolite is also consistent with low-pressure hornblende-granulite grade metamorphic conditions (Miyashiro, 1973).

### PETROLOGY OF GARNET-CORDIERITE GNEISSES

#### *Petrography*

Locations of the six kinzigite samples examined in this study are shown in Fig. 2. All of the samples contain K-feldspar + plagioclase + quartz + biotite + garnet + cordierite + sillimanite + ilmenite + zircon + hercynite + graphite.

Cordierite occurs as elongate crystals parallel to biotite and together with sillimanite defines a single well-developed foliation (Fig. 3A). They are optically negative with a  $2V$  of  $\sim 70$ – $80^\circ$ , indicating less than 1 wt.% channel-filling molecular  $H_2O$  or  $CO_2$  (Armbruster & Bloss, 1982). Cordierite generally occurs in close spatial association with biotite but is also found in contact with feldspars and garnet.

Two textural varieties of sillimanite are present. One variety consists of well-developed acicular crystals with easily discernible basal parting occurring predominantly within cordierite. This sillimanite parallels the trace of the foliation defined by both biotite and cordierite in thin section. The second variety of sillimanite consists of mats of fine-grained fibrolite within and between crystals of cordierite, and to a lesser extent, biotite. Where in association with cordierite, fine-grained fibrolite generally extends from grain boundaries into the host cordierite at approximately right angles to the host crystal edges. The occurrence of fibrolite in and around cordierite is most prominent near contacts with K-feldspar.

Two textural varieties of garnet can also be distinguished. Subhedral poikiloblasts enclosing quartz and biotite are volumetrically dominant, but garnet also occurs as anhedral intergrowths with biotite (Fig. 3B). These intergrowths are generally parallel to cleavage traces in biotite and to the foliation. In some cases, the intergrowths extend outwards from subhedral crystals into touching biotites.

K-feldspar forms large (up to 20 mm) perthitic poikiloblasts enclosing subhedral garnet, biotite, quartz, plagioclase, and, very rarely, cordierite and its inclusions. It also occurs as interstitial anhedral with well-developed microcline twinning. Within the large poikiloblasts, biotites are generally oriented parallel to the trace of the foliation.

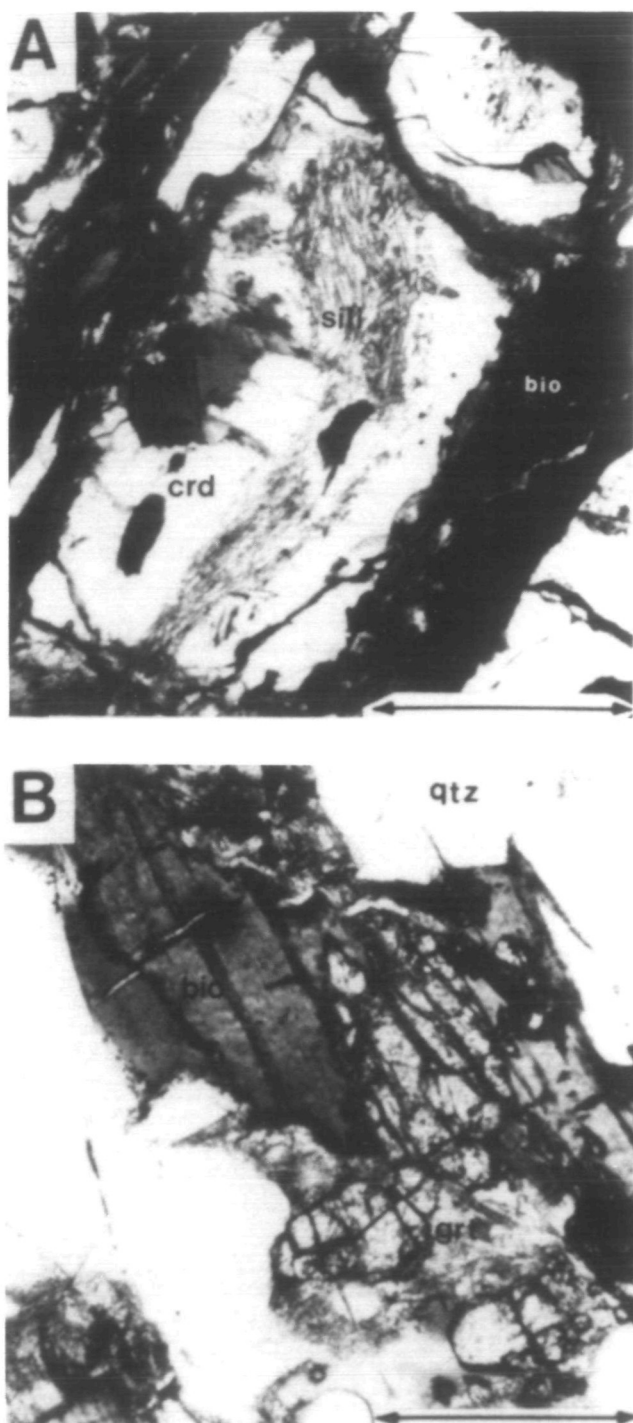


FIG. 3. Photomicrographs of garnet-cordierite gneiss. (A) Cordierite crystals elongate parallel to surrounding biotite and sillimanite. (B) Garnet intergrown with biotite. Abbreviations: bio, biotite; crd, cordierite; grt, garnet; sill, sillimanite; and qtz, quartz. Bars are 1 mm in length.

Subhedral plagioclase is ubiquitous and myrmekite is common. Ilmenite occurs as a matrix phase and as inclusions in cordierite and biotite. Hercynite is found only poikilitically enclosed in cordierite. Alteration is limited to minor sericitization of plagioclase and development of chlorite after biotite.

The textures described above are indicative of a single metamorphic event. Several characteristics of this event, as expressed in the garnet-cordierite gneisses, can be gleaned from the petrographic observations. Biotite and cordierite grew under conditions of deviatoric stress. Garnet apparently formed in two stages: an initial growth of subhedral equant porphyroblasts of garnet enveloping biotite, and a later growth of anhedral garnet apparently reacting with biotite. Nucleation of K-feldspar may also have occurred in two distinct stages, represented by interstitial microcline and large poikiloblastic perthite. K-feldspar poikiloblasts enclose all major phases, but cordierite, sillimanite, and hercynite are exceedingly rare as inclusions. Poikiloblastic growth of K-feldspar probably occurred during or in part prior to growth of these phases. Mats of fibrolite are interpreted to represent retrograde growth of sillimanite. Support for this interpretation includes the occurrence of fibrolite perpendicular to grain boundaries and textures indicating growth at the expense of cordierite, biotite, and K-feldspar.

The origin of the coarser grained sillimanite included in cordierite is problematic. Growth parallel to foliation is suggestive of a prograde nucleation. This interpretation suggests

TABLE 1  
*Electron microprobe analyses of representative biotites*

| Sample                         | MC-85-14<br>1c | MC-85-14<br>4 | MC-85-23<br>1 | MC-85-23<br>2 | FL-003<br>1 | FL-003<br>2 | FL-001<br>1 | FL-001<br>2 |
|--------------------------------|----------------|---------------|---------------|---------------|-------------|-------------|-------------|-------------|
| <i>Wt.% oxides</i>             |                |               |               |               |             |             |             |             |
| SiO <sub>2</sub>               | 37.28          | 37.20         | 35.12         | 35.03         | 36.42       | 35.15       | 35.83       | 35.71       |
| Al <sub>2</sub> O <sub>3</sub> | 17.18          | 17.05         | 17.01         | 16.97         | 17.15       | 17.44       | 17.64       | 17.02       |
| TiO <sub>2</sub>               | 3.31           | 4.32          | 3.50          | 3.70          | 3.06        | 3.85        | 3.56        | 3.35        |
| FeO*                           | 17.19          | 17.28         | 22.58         | 22.38         | 19.47       | 20.09       | 16.81       | 19.14       |
| MnO                            | 0.02           | 0.05          | 0.08          | 0.10          | 0.00        | 0.01        | 0.01        | 0.04        |
| MgO                            | 11.36          | 10.73         | 7.88          | 7.62          | 11.22       | 9.79        | 11.84       | 9.58        |
| CaO                            | 0.00           | 0.00          | 0.05          | 0.00          | 0.01        | 0.02        | 0.01        | 0.03        |
| Na <sub>2</sub> O              | 0.07           | 0.08          | 0.08          | 0.06          | 0.14        | 0.12        | 0.07        | 0.02        |
| K <sub>2</sub> O               | 9.69           | 9.79          | 9.57          | 9.60          | 9.54        | 9.86        | 9.47        | 9.83        |
| Total                          | 96.11          | 96.52         | 95.89         | 95.48         | 97.03       | 96.36       | 95.28       | 94.76       |
| <i>Cations per 11 oxygens</i>  |                |               |               |               |             |             |             |             |
| Si                             | 2.780          | 2.767         | 2.711         | 2.714         | 2.725       | 2.671       | 2.698       | 2.742       |
| Al                             | 1.510          | 1.495         | 1.547         | 1.549         | 1.512       | 1.562       | 1.566       | 1.541       |
| Al <sup>VI</sup>               | 0.290          | 0.262         | 0.258         | 0.263         | 0.237       | 0.233       | 0.264       | 0.283       |
| Al <sup>IV</sup>               | 1.220          | 1.233         | 1.289         | 1.286         | 1.275       | 1.329       | 1.302       | 1.258       |
| Ti                             | 0.186          | 0.242         | 0.203         | 0.216         | 0.172       | 0.220       | 0.202       | 0.194       |
| Fe <sup>2+*</sup>              | 1.072          | 1.075         | 1.457         | 1.450         | 1.218       | 1.277       | 1.059       | 1.229       |
| Mn                             | 0.002          | 0.003         | 0.005         | 0.007         | 0.000       | 0.001       | 0.001       | 0.003       |
| Mg                             | 1.262          | 1.190         | 0.907         | 0.880         | 1.252       | 1.109       | 1.330       | 1.098       |
| Ca                             | 0.000          | 0.000         | 0.004         | 0.000         | 0.001       | 0.002       | 0.002       | 0.002       |
| Na                             | 0.011          | 0.013         | 0.013         | 0.010         | 0.021       | 0.018       | 0.010       | 0.004       |
| K                              | 0.922          | 0.929         | 0.942         | 0.949         | 0.911       | 0.956       | 0.910       | 0.964       |
| Total                          | 7.745          | 7.715         | 7.790         | 7.775         | 7.812       | 7.815       | 7.778       | 7.777       |

\* Total Fe.

cordierite grew at the expense of biotite and sillimanite, as implied by the occurrence of cordierite almost exclusively within accumulations of biotite, separation of biotite and sillimanite by intervening cordierite, and lack of evidence for cordierite formation by garnet breakdown (the other major Fe–Mg aluminosilicate). However, epitaxial growth of sillimanite at the expense of cordierite cannot be ruled out on the basis of textural evidence. Cho & Fawcett (1986) have shown that cordierite may grow by coalescence of pseudohexagonal crystals resulting in a crystal substrate with channels parallel to the *c*-axis. These channels could provide suitable sites for growth of acicular sillimanite parallel to the long dimension of cordierite host crystals.

### Mineral chemistry

#### Biotite

Table 1 shows representative microprobe analyses of biotite from several of the rock samples examined. The compositions of biotite from all six samples are shown in the ideal biotite plane annite–phlogopite–siderophyllite–eastonite in Fig. 4. For comparison, the data compiled by Guidotti (1984) were used to define fields for biotite from pelites of granulite grade with the assemblage cordierite  $\pm$  aluminosilicate  $\pm$  garnet, and biotites from upper amphibolite to granulite grade pelites with K-feldspar + aluminosilicate and no muscovite (Fig. 4). The majority of analyses fall within the cordierite  $\pm$  aluminosilicate  $\pm$  garnet field. Biotites poikilolitically enclosed in cordierite have anomalously high  $\text{Al}^{\text{VI}}$ . These biotites are also distinctive in their relatively low Ti content (0.074–0.117 atoms/11 oxygens, cf. Table 1) and high  $\text{Mg}/(\text{Mg} + \text{Fe})$ . The significance of these anomalous biotite compositions is discussed in the analysis of the cordierite-forming reaction in a separate paper (Part II, Young, 1989).

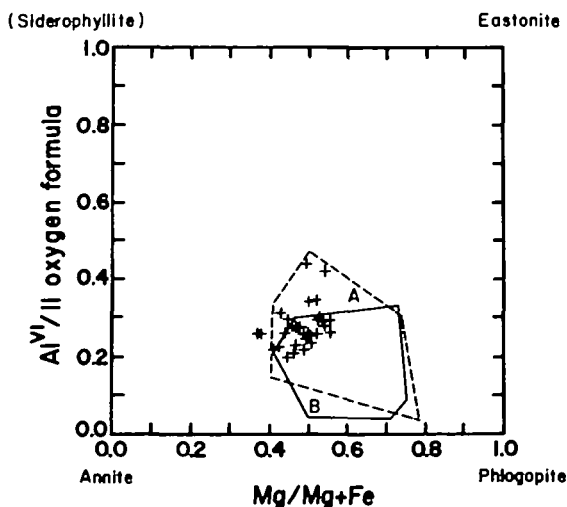
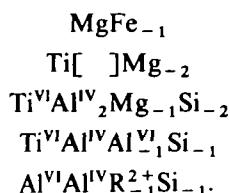


FIG. 4. Compositions of representative biotites from the McCullough pelites in the ideal biotite plane annite–phlogopite–siderophyllite–eastonite. Fields A (dashed line) and B (solid line) are defined by the data compiled by Guidotti (1984). Field A represents biotites from upper amphibolite to granulite grade pelites with K-feldspar + aluminosilicate and no muscovite. Field B represents biotites coexisting with cordierite  $\pm$  aluminosilicate  $\pm$  garnet in granulite-grade pelites.

Operative substitutions as indicated by bivariate plots and principal components analysis are, in approximate order of decreasing importance,



The negative correlation between  $\text{Ti}^{\text{VI}}$  and Mg may in part have been crystallographically controlled by affording a reduction in the misfit between octahedral and tetrahedral layers (Guidotti, 1984). The  $\text{Ti}[\text{ }]\text{Mg}_{-2}$  substitution has been shown to be important in other aluminum and titanium saturated rocks of granulite grade (Dymek, 1983; Guidotti, 1984; Indares & Martignole, 1985; see also Otten & Buseck, 1987).

### Garnet

Garnet of the gneisses is iron rich with molar end-member proportions approximating 80% almandine, 14% pyrope, 4% spessartine, and 2% grossular. Representative analyses are shown in Table 2. *t*-tests for the variables Mg and Fe per 12 oxygens indicate that sample means for all rock samples are not representative of the means for analyses from any one rock at the 90% confidence level. The results are consistent with the biotite data in showing significantly greater inter-rock variation in mineral chemistry relative to intra-rock variation.

Zoning in the garnets is characterized by strong compositional changes near poikilitically enclosed biotite and relatively minor variations away from biotite inclusions (Fig. 5). Ca contents are approximately constant. Mn concentrations are nearly constant, with a minor

TABLE 2  
*Electron microprobe analyses of representative garnets*

| Sample                         | MC-85-14<br>1r | MC-85-14<br>4c | MC-85-14<br>6c | FL-003<br>1-1 | FL-003<br>1-4 | FL-003<br>2r | FL-001<br>1-1 |
|--------------------------------|----------------|----------------|----------------|---------------|---------------|--------------|---------------|
| <i>Wt.% oxides</i>             |                |                |                |               |               |              |               |
| SiO <sub>2</sub>               | 39.40          | 38.95          | 38.96          | 39.27         | 39.29         | 38.86        | 38.75         |
| Al <sub>2</sub> O <sub>3</sub> | 18.99          | 19.95          | 20.27          | 19.67         | 19.95         | 19.92        | 19.78         |
| TiO <sub>2</sub>               | 0.01           | 0.01           | 0.23           | 0.00          | 0.00          | 0.00         | 0.03          |
| FeO*                           | 33.25          | 32.97          | 33.05          | 36.61         | 36.17         | 35.16        | 34.81         |
| MnO                            | 2.25           | 2.35           | 2.47           | 1.74          | 1.79          | 1.74         | 0.52          |
| MgO                            | 4.26           | 4.32           | 4.57           | 3.52          | 3.43          | 3.95         | 4.51          |
| CaO                            | 0.76           | 0.86           | 0.81           | 0.95          | 0.99          | 0.98         | 0.97          |
| Total                          | 98.95          | 99.45          | 100.16         | 101.78        | 101.64        | 100.61       | 99.37         |
| <i>Cations Per 12 oxygens</i>  |                |                |                |               |               |              |               |
| Si                             | 3.167          | 3.113          | 3.090          | 3.106         | 3.106         | 3.092        | 3.103         |
| Al                             | 1.799          | 1.880          | 1.896          | 1.834         | 1.859         | 1.869        | 1.867         |
| Ti                             | 0.001          | 0.001          | 0.002          | 0.000         | 0.000         | 0.000        | 0.002         |
| Fe <sup>2+*</sup>              | 2.236          | 2.203          | 2.193          | 2.422         | 2.391         | 2.340        | 2.331         |
| Mn                             | 0.153          | 0.160          | 0.166          | 0.117         | 0.120         | 0.118        | 0.036         |
| Mg                             | 0.511          | 0.515          | 0.541          | 0.416         | 0.405         | 0.469        | 0.538         |
| Ca                             | 0.066          | 0.074          | 0.069          | 0.081         | 0.084         | 0.084        | 0.084         |
| Total                          | 7.933          | 7.946          | 7.963          | 7.977         | 7.964         | 7.972        | 7.961         |

\* Total Fe.

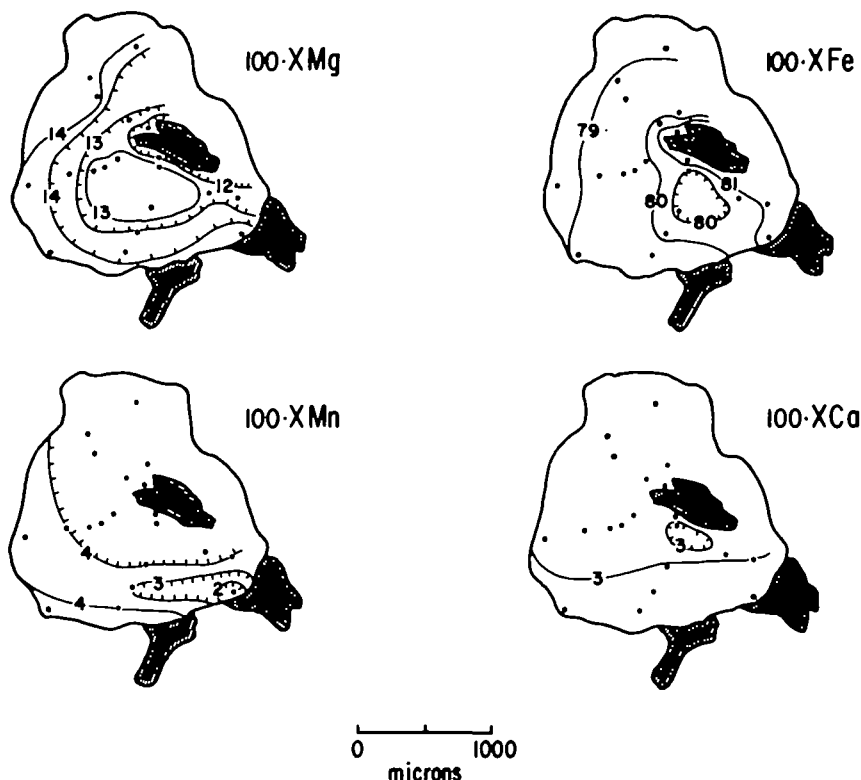


FIG. 5. Composition contour maps for a garnet poikilolithically enclosing biotite (lined pattern) from sample FL-003. Note abrupt changes in  $Mg/(Mg + Fe + Ca + Mn)$  and  $Fe/(Mg + Fe + Ca + Mn)$  near the included biotite crystal. Dots indicate microprobe analysis points.

negative correlation between Mn and Mg indicated in the interiors of a few grains. In garnets lacking inclusions, Mg/Fe is highest in the cores. In grains containing biotite inclusions, Mg/Fe decreases abruptly near the inclusions. Similar characteristics have been observed in other garnets from granulite grade pelites (Tracy *et al.*, 1976; Grew, 1981; Tracy & Dietsch, 1982; Indares & Martignole, 1984; Schreurs & Westra, 1986) and have been interpreted as the result of extensive homogenization by cation volume diffusion at temperatures exceeding 600 °C (Woodsworth, 1977) followed by local exchange of Fe and Mg between garnet and biotite with cooling (Tracy *et al.*, 1976). No significant compositional discontinuity has been observed at grain edges (< 50  $\mu m$  e.g., Hodges & Royden, 1984).

#### Cordierite

Cordierite compositions yield a mean  $Fe/(Fe + Mg)$  of 0.411. Although significant differences between analyses from different rock samples are evident, *t*-tests for the variables Fe and Mg suggest a single sample mean is adequate for the combined data. Elevated oxide totals are characteristic of the electron microprobe analyses (Table 3) and are consistent with the optical properties in suggesting low total volatile content (0.00–0.42 moles  $H_2O$  equivalent per 18 oxygen formula unit) in cordierite channels. The negative optic sign indicates low  $CO_2$  content in the channel filling fluid. Na content is low, with total alkalis rarely exceeding 0.20 wt. %.

TABLE 3  
Electron microprobe analyses of representative cordierites

| Sample                         | MC-85-14<br>1r | MC-85-23<br>1r | MC-85-23<br>2c | FL-003<br>1r | FL-003<br>3-1Rc | FL-003<br>3-3Rr |
|--------------------------------|----------------|----------------|----------------|--------------|-----------------|-----------------|
| <i>Wt.% oxides</i>             |                |                |                |              |                 |                 |
| SiO <sub>2</sub>               | 49.44          | 48.34          | 48.45          | 50.02        | 49.33           | 48.66           |
| Al <sub>2</sub> O <sub>3</sub> | 32.91          | 33.36          | 32.60          | 33.61        | 33.59           | 33.02           |
| TiO <sub>2</sub>               | 0.04           | 0.04           | 0.00           | 0.00         | 0.01            | 0.03            |
| FeO*                           | 7.86           | 11.03          | 11.57          | 9.55         | 7.87            | 9.78            |
| MnO                            | 0.15           | 0.19           | 0.21           | 0.03         | 0.07            | 0.21            |
| MgO                            | 8.27           | 6.58           | 6.48           | 7.94         | 9.34            | 7.79            |
| CaO                            | 0.00           | 0.01           | 0.00           | 0.00         | 0.09            | 0.09            |
| Na <sub>2</sub> O              | 0.05           | 0.06           | 0.10           | 0.05         | 0.04            | 0.10            |
| Total                          | 98.74          | 99.61          | 99.43          | 101.21       | 100.34          | 99.68           |
| <i>Cations per 18 oxygens</i>  |                |                |                |              |                 |                 |
| Si                             | 5.050          | 4.972          | 5.007          | 5.020        | 4.965           | 4.966           |
| Al                             | 3.962          | 4.044          | 3.971          | 3.975        | 3.986           | 4.000           |
| Ti                             | 0.003          | 0.003          | 0.000          | 0.000        | 0.000           | 0.003           |
| Fe <sup>2+*</sup>              | 0.672          | 0.949          | 1.000          | 0.802        | 0.662           | 0.834           |
| Mn                             | 0.013          | 0.017          | 0.018          | 0.003        | 0.006           | 0.018           |
| Mg                             | 1.259          | 1.009          | 0.999          | 1.188        | 1.401           | 1.185           |
| Ca                             | 0.000          | 0.002          | 0.001          | 0.000        | 0.009           | 0.009           |
| Na                             | 0.011          | 0.012          | 0.021          | 0.010        | 0.008           | 0.020           |
| Total                          | 10.979         | 11.008         | 11.017         | 10.998       | 11.037          | 11.035          |

\* Total Fe.

Zoning in cordierites is highly irregular. The irregularity reflects variable dominance of the effects of small scale cation exchange with inclusions, diffusion, and continuous reequilibration with matrix grains following nucleation. Notably, Mg/Fe in cordierite is dramatically elevated adjacent to enclosed garnets and biotites (Part II).

### Feldspars

Plagioclase compositions yield 0.32 average mol fraction anorthite ( $X_{An}$ ) with a range of 0.34–0.29. Zoning is weak, with an average core to rim variation of  $\pm 0.03 X_{An}$ . K-feldspar analyses are more variable in part because of exsolution. Integrated compositions range from 0.68 to 0.86 mol fraction orthoclase ( $X_{Or}$ ) with a majority of point analyses falling within the range 0.80–0.85  $X_{Or}$  (Table 4).

### Oxides

Microprobe analyses and energy dispersive spectra indicate that the opaque phases are nearly end-member ilmenite with less than 0.3 mol fraction MnTiO<sub>3</sub> (pyrophanite component) and 0.2 mol fraction Fe<sub>2</sub>O<sub>3</sub> (hematite component).

Green spinels are hercynite–spinel solid solutions with between 0.83 and 0.95 mol fraction hercynite ( $X_{Hr}$ ) component. A single analysis for Zn (Table 5) yielded 4.36 wt.% ZnO. Low totals for analyses with Zn omitted are consistent with this concentration.

### Mineral equilibria

Compositions of coexisting biotite, garnet, and cordierite grains within 3 mm of each other yield a limited range in  $K_B^{(Mg/Fe)}$ , with covariation of (Mg/Fe) for all three minerals

TABLE 4  
*Electron microprobe analyses of representative feldspars*

| Sample                           | Plagioclase    |                |              |              | K-feldspar  |                |               |
|----------------------------------|----------------|----------------|--------------|--------------|-------------|----------------|---------------|
|                                  | MC-85-14<br>1r | MC-85-23<br>2r | FL-003<br>1c | FL-003<br>2r | FL-001<br>1 | *MC-85-14<br>1 | MC-85-23<br>2 |
| Wt.% oxides                      |                |                |              |              |             |                |               |
| SiO <sub>2</sub>                 | 62.26          | 61.34          | 60.83        | 59.36        | 60.07       | 64.52          | 64.96         |
| Al <sub>2</sub> O <sub>3</sub>   | 23.66          | 25.25          | 25.27        | 25.37        | 24.64       | 19.86          | 19.30         |
| TiO <sub>2</sub>                 | 0.00           | 0.02           | 0.00         | 0.02         | 0.01        | 0.05           | 0.07          |
| Fe <sub>2</sub> O <sub>3</sub> † | 0.06           | 0.02           | 0.00         | 0.04         | 0.00        | 0.04           | 0.06          |
| MnO                              | 0.00           | 0.01           | 0.02         | 0.00         | 0.00        | 0.00           | 0.00          |
| MgO                              | 0.02           | 0.01           | 0.00         | 0.00         | 0.00        | 0.02           | 0.00          |
| CaO                              | 5.11           | 6.13           | 6.27         | 6.52         | 5.57        | 1.06           | 0.13          |
| Na <sub>2</sub> O                | 8.77           | 7.69           | 7.72         | 7.83         | 7.89        | 3.78           | 1.63          |
| K <sub>2</sub> O                 | 0.17           | 0.14           | 0.19         | 0.14         | 0.13        | 10.01          | 14.11         |
| Total                            | 100.05         | 100.66         | 100.32       | 99.28        | 98.36       | 99.34          | 100.27        |
| Cations per 8 oxygens            |                |                |              |              |             |                |               |
| Si                               | 2.759          | 2.704          | 2.693        | 2.663        | 2.708       | 2.941          | 2.971         |
| Al                               | 1.236          | 1.312          | 1.319        | 1.342        | 1.309       | 1.067          | 1.040         |
| Ti                               | 0.000          | 0.001          | 0.000        | 0.001        | 0.001       | 0.002          | 0.002         |
| Fe <sup>3+</sup> †               | 0.002          | 0.001          | 0.000        | 0.001        | 0.000       | 0.002          | 0.002         |
| Mn                               | 0.000          | 0.001          | 0.001        | 0.000        | 0.000       | 0.000          | 0.000         |
| Mg                               | 0.001          | 0.001          | 0.000        | 0.000        | 0.000       | 0.001          | 0.000         |
| Ca                               | 0.243          | 0.289          | 0.297        | 0.313        | 0.269       | 0.052          | 0.007         |
| Na                               | 0.754          | 0.657          | 0.663        | 0.681        | 0.690       | 0.334          | 0.145         |
| K                                | 0.010          | 0.008          | 0.011        | 0.008        | 0.008       | 0.582          | 0.823         |
| Total                            | 5.005          | 4.972          | 4.984        | 5.009        | 4.986       | 4.981          | 4.990         |

\* Integrated.

† Total Fe.

TABLE 5  
*Electron microprobe analyses of representative oxides*

| Sample                         | Hercynite   |                              | Ilmenite    |
|--------------------------------|-------------|------------------------------|-------------|
|                                | FL-003<br>2 | FL-003<br>1                  | FL-003<br>1 |
| <i>Wt.%</i>                    |             |                              |             |
| FeO*                           | 33.68       | 35.06                        | 45.37       |
| TiO <sub>2</sub>               | 0.00        | 0.03                         | 49.55       |
| Al <sub>2</sub> O <sub>3</sub> | 56.84       | 54.97                        | 0.00        |
| MgO                            | 2.84        | 2.71                         | 0.03        |
| ZnO                            | 4.36        | n.a.                         | n.a.        |
| MnO                            | n.a.        | 0.16                         | 1.00        |
| Total                          | 97.72       | 93.46                        | 96.09       |
| <i>Cations per 4 oxygens</i>   |             | <i>Cations per 3 oxygens</i> |             |
| Fe <sup>2+</sup> *             | 0.828       | 0.893                        | 1.002       |
| Ti                             | 0.000       | 0.001                        | 0.984       |
| Al                             | 1.969       | 1.974                        | 0.000       |
| Mg                             | 0.124       | 0.123                        | 0.001       |
| Zn                             | 0.095       | n.a.                         | n.a.        |
| Mn                             | n.a.        | 0.004                        | 0.022       |
| Total                          | 3.016       | 3.024                        | 2.013       |

\*. Total Fe.

n.a. = not analyzed.

(Fig. 6).  $K_D^{(\text{Mg/Fe})}$  for biotite and garnet (Fig. 6) varies from 3.15 to 5.99 with most points lying near  $K_D = 5.0$ . In general, core compositions for both biotite and garnet yield higher  $K_D$ s within any one sample. The  $K_D^{(\text{Mg/Fe})}$  values for cordierite and garnet (Fig. 6) range from 6.16 to 10.70. Core compositions for garnet and cordierite yield generally lower  $K_D$ s. Cordierite–biotite  $K_D^{(\text{Mg/Fe})}$  varies from 1.13 to 1.82 (Fig. 6) with most pairs falling within the range 1.42 to 1.82. The lowest distribution coefficients are from points derived from inclusions of biotite in cordierite.

AFM assemblages from individual samples yield uniform tie line slopes between rim compositions within 3 mm domains. Crossing tie lines exist between the three-phase assemblages biotite + hercynite + cordierite and biotite + garnet + cordierite (Fig. 7). Consistency of partitioning among phases within 3 mm domains establishes the scale of mosaic equilibrium. Crossing of the tie lines hercynite–biotite and garnet–cordierite is suggestive of a reaction relationship and is consistent with the occurrence of hercynite as inclusions in cordierite. Similarly, crossing of the biotite–sillimanite and garnet–cordierite tie lines suggests that biotite + sillimanite stability gave way to garnet + cordierite.

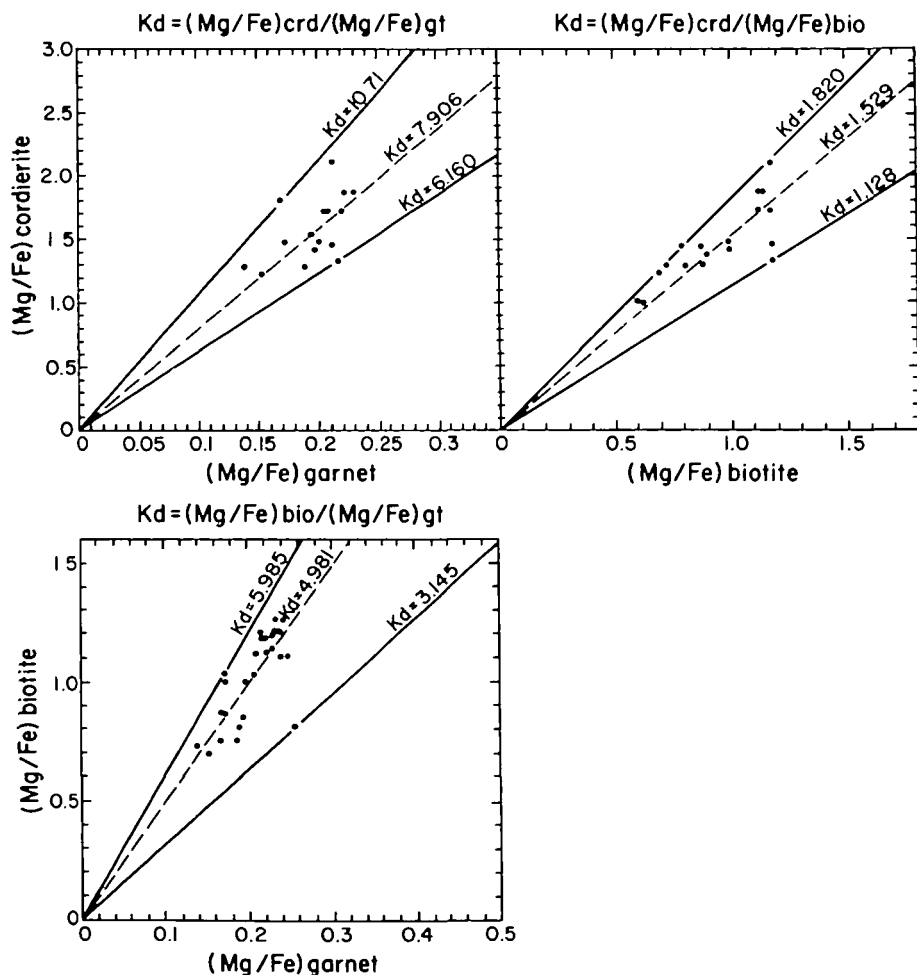


FIG. 6. Atomic ratios (Mg/Fe) for coexisting garnet, biotite, and cordierite

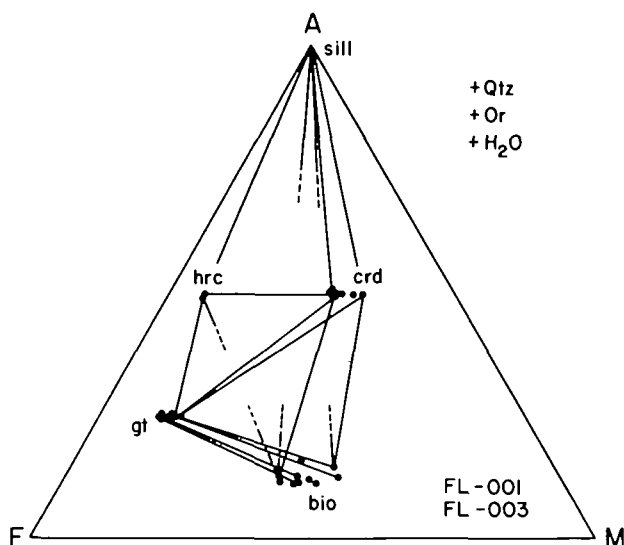


FIG. 7.  $A(\text{Al}_2\text{O}_3)$ – $F(\text{FeO})$ – $M(\text{MgO})$  molecular proportions projected from quartz, orthoclase, and  $\text{H}_2\text{O}$ . Tie lines connecting phases are partially omitted for clarity.

The system  $\text{K}_2\text{O}$ – $\text{TiO}_2$ – $\text{Na}_2\text{O}$ – $\text{CaO}$ – $\text{FeO}$ – $\text{MgO}$ – $\text{MnO}$ – $\text{Al}_2\text{O}_3$ – $\text{SiO}_2$ – $\text{H}_2\text{O}$  (KTNCFMMASH) permits hercynite + garnet + cordierite + biotite + sillimanite + plagioclase + K-feldspar + quartz + ilmenite to exist as a trivariant assemblage. However, equilibrium between hercynite and other phases, with the exception of cordierite, is not texturally demonstrable.

#### Thermobarometry

The minerals in textural equilibrium in the garnet–cordierite gneisses permit application of several geothermobarometers. Composition pairs used include rims (within  $\sim 200\ \mu\text{m}$  of the grain edge) of grains that are either touching or are in close proximity ( $< 1\ \text{mm}$ ), comprising individual domains in a system of mosaic equilibrium.

The garnet–biotite thermometer calibrations of Goldman & Albee (1977) and Ferry & Spear (1978), and the reformulations of Ganguly & Saxena (1984), Hodges & Royden (1984), and Indares & Martignole (1985) have been used to calculate temperatures for the six samples analyzed. A typical result for a mineral domain from sample FL-003 is illustrated in Fig. 8. Formulations of Ferry & Spear (1978), Ganguly & Saxena (1984), and Hodges & Royden (1984) yield similar results within reasonable estimates for minimum uncertainty (e.g., Hodges & McKenna, 1987). Calculated temperatures range from 540 to  $700^\circ\text{C}$  depending upon pressure and composition pair, with a mode at  $\sim 600^\circ\text{C}$  (Fig. 9). The calibration of Goldman & Albee (1977) yields systematically lower temperatures, apparently in part because their calibration points lie well below  $600^\circ\text{C}$  (Fig. 8).

The revision of Indares and Martignole (1985) for garnet–biotite Fe–Mg exchange applies an empirically derived correction factor for the  $\text{Ti}^{\text{VI}}[\text{ }^{\text{VI}}\text{R}_{-2}^{2+}]$  substitution in biotite to the experimental data of Ferry & Spear (1978). This substitution, as noted above, was operative in the McCullough biotites. The correction factor for  $\text{Al}^{\text{VI}}$  substitution, also used in the calibration, is an order of magnitude smaller than the Ti factor. Indares & Martignole derived the corrections by regression using the differences between observed equilibrium

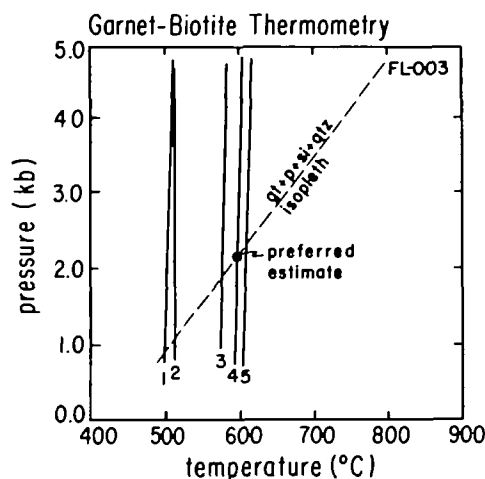


FIG. 8. Comparison of results from garnet-biotite thermometer calibrations of Indares & Martignole (1985) (1), Goldman & Albee (1977) (2), Ganguly & Saxena (1984) (3), Ferry & Spear (1978) (4), and Hodges & Royden (1984) (5) for a single mineral pair from sample FL-003. The corresponding isopleth for garnet (gt) + plagioclase (p) + sillimanite (si) + quartz (qtz) equilibrium using the calibration of Ghent *et al.* (1979) is also shown. The choice of the preferred calibration is discussed in the text.

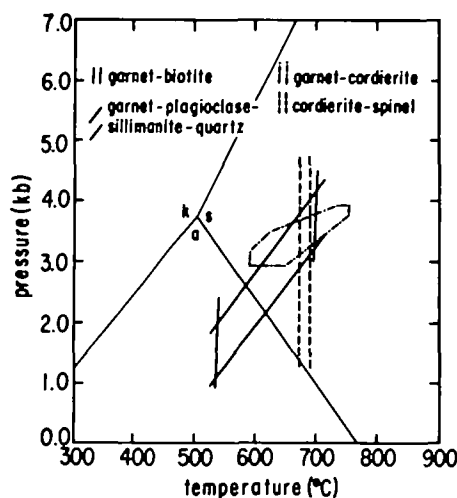


FIG. 9. Summary of calculated pressures and temperatures for the McCullough pelites using garnet-biotite (Ferry & Spear, 1978), garnet-plagioclase-sillimanite-quartz (Ghent *et al.*, 1979), garnet-cordierite (Holdaway & Lee, 1977; Martignole & Sisi, 1981; Aranovich & Podlesskii, 1983), and cordierite-spinel (Vielzeuf, 1983) exchange equilibria. The garnet-cordierite field encloses the range in calculated  $P$ - $T$  points derived using that equilibrium. The other lines are limited for calculated isopleths (constant  $K_D$ ) for the indicated equilibria. The aluminium silicate phase diagram of Holdaway (1971) is shown for reference.

constants ( $K_{eq}$ ) at  $T$  and  $P$  for the Fe-Mg exchange reaction and the product  $K_D \cdot (\gamma Fe / \gamma Mg)^{garnet}$ , where  $K_D$  is the distribution coefficient and  $\gamma Fe$  and  $\gamma Mg$  are activity coefficients for these elements in the garnets used in the reformulation. The relatively low  $X_{Ca}$  in the McCullough pelite garnets indicates that the difference between  $K_{eq}$  and  $K_D$  for the garnet-biotite Fe-Mg exchange reaction in these rocks is smaller than the differences

used by Indares & Martignole in their regression. The more ideal behavior of the garnet Fe–Mg mixing relative to that in the garnets used to derive the correction factors suggests that the factors are too large for the McCullough pelites (Indares & Martignole, 1985, p. 277), resulting in temperature estimates lower than those from the other formulations (Fig. 8). The overcorrection of the Indares & Martignole (1985) thermometer and unreliability of the Goldman & Albee (1977) calibration at high grades leads us to favor temperatures calculated using the Ferry & Spear (1978) calibration for garnet–biotite exchange (consistent with those using the Ganguly & Saxena and Hodges & Royden formulations).

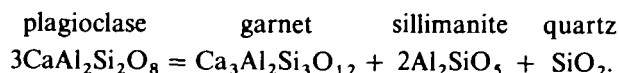
Fe–Mg exchange between cordierite and garnet permits another estimate of temperature, and can also constrain the pressures of metamorphism. The calibrations of Holdaway & Lee (1977), Martignole & Sisi (1981), and Aranovich & Podlesskii (1983) have been applied. A primary difference in the calibrations is the treatment of the thermodynamic effects of channel filling  $H_2O$  in cordierite, as discussed by Wood (1973). As noted above, the optical properties and high totals from microprobe analyses suggest low volatile content in the cordierites of the McCullough pelites. Thus, the  $Mg/(Mg + Fe)$  isopleths for garnet and cordierite with  $nH_2O$  (moles  $H_2O$  in cordierite per 18 oxygen formula unit) = 0 proposed by Martignole & Sisi (1981) were used for constraining  $P$  and  $T$ .

Temperatures calculated using the Holdaway & Lee (1977) calibration for garnet–cordierite Fe–Mg exchange range from 590 to 720°C depending on paired compositions and pressure. These temperatures are within the high temperature portion of the range defined by garnet–biotite thermometry (Fig. 9). The isopleth diagram of Martignole & Sisi (1981) for garnet and cordierite yields a temperature range of 590–750°C in agreement with that derived from the Holdaway & Lee calibration, and a pressure range of 2.9–3.9 kb (Fig. 9). The effect of assuming  $nH_2O = 0.50$  (higher than the maximum for the McCullough pelites) is to increase the estimated pressures of equilibration by approximately 700 b. Pressure–temperature estimates obtained from the experimental calibration of Aranovich & Podlesskii (1983) coincide with those derived from the theoretical model of Martignole & Sisi (1981) using the  $H_2O$ -poor fluid equation of the former.

Cordierite composition alone can be used to constrain possible  $T$  and  $P$  using the Martignole & Sisi isopleths. These estimates do not require specification of garnet–cordierite equilibrium pairs. The observed cordierite compositions permit a maximum  $T$  of 790°C at pressures  $\leq 4$  kb and a pressure range of 2.2–4.9 kb from 600–800°C respectively.

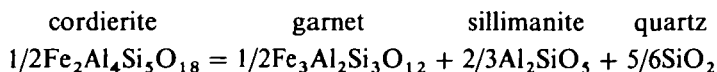
Vielzeuf (1983) suggested a pressure independent calibration of the temperature sensitivity of  $Fe^{2+}$ –Mg exchange between cordierite and a hercynite–spinel solid solution phase. Temperatures calculated using this calibration range from 673–692°C and are similar to results from the other exchange equilibria (Fig. 9).

Additional pressure constraints were obtained using the calibration of Ghent *et al.* (1979) for the equilibrium:



The range of calculated isopleths for this equilibrium is shown in Fig. 9. The intersections of the garnet–plagioclase equilibrium isopleths with the preferred garnet–biotite Fe–Mg distribution isopleths define a pressure–temperature window with extrema at 535°C at 1.1 kb and 700°C at 4.2 kb (Fig. 9) and a clustering near the median of 615°C at 2.7 kb. The  $P$ – $T$  constraints indicated by garnet–cordierite equilibria fall within the upper  $P$  and  $T$

portion of this window as do temperatures calculated using cordierite-spinel Fe-Mg distribution. The geobarometer of Bhattacharya (1986) for the equilibrium:



yields pressures approximately 1 kb higher than the calibrations of Martignole & Sisi (1981) and Aranovich & Podlesskii (1983) using various garnet activity formulations.

In these thermobarometry calculations,  $\text{Fe}^{3+}$  has not been considered, the assumption being that total  $\text{Fe} \cong \text{Fe}^{2+}$ . The calculated garnet stoichiometries do not permit significant andradite component. However,  $\text{Fe}^{3+}$  in biotite cannot be estimated quantitatively with available data. The Mössbauer spectral study of naturally occurring biotites by Dyar & Burns (1986) confirms the importance of  $\text{Fe}^{3+}$  in biotite chemistry. Significant unrecognized  $\text{Fe}^{3+}$  in biotite, not compensated by  $\text{Fe}^{3+}$  in garnet, would result in erroneously high temperatures using the calibrations for garnet-biotite  $\text{Fe}^{2+}$ -Mg distribution. Ghent *et al.* (1979) noted that ignoring andradite in garnet typically results in an overestimate of pressure amounting to less than 300 b. However, Bohlen *et al.* (1983) noted that at low grossular contents, such as that of the McCullough pelite garnets, unconstrained non-ideal mixing may hamper the reliability of the plagioclase-garnet- $\text{Al}_2\text{SiO}_5$ -quartz barometer.

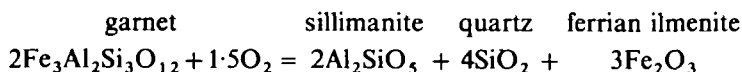
The lowest temperature recorded by garnet-cordierite Fe-Mg partitioning, 590°C at 3.0 kb, is derived from nearby points on a small garnet enclosed by cordierite. The garnet touches biotite also enclosed in cordierite. The garnet-biotite pair yields a temperature of 575°C at 3.0 kb. These results suggest that cation exchange between cordierite, garnet, and biotite in close spatial association attained closure with respect to  $T$  (and  $P$ ). A similar closure effect between these phases has been described by Tracy & Dietsch (1982). This closure and the similarly consistent conditions recorded by matrix-assemblage cordierite, garnet, and biotite suggest that the magnitude of systematic errors between calibrations is less than the actual differences in  $T$  and  $P$  preserved in different local domains.

#### $a_{\text{H}_2\text{O}}$ and $f_{\text{O}_2}$

Equilibrium among biotite, sillimanite, quartz, garnet, and K-feldspar was used to calculate the activity of  $\text{H}_2\text{O}$  in the paragneiss samples relative to a standard state defined as pure  $\text{H}_2\text{O}$  vapor at  $P$  and  $T$  using the method of Phillips (1980). Calculated activities derived from rim compositions range from 0.03 to 0.26 with a mean of 0.16. These results are insensitive to the form of the ideal contributions to the activity expressions.

The occurrence of nearly end-member ilmenite ( $X_{\text{ilm}} = 0.92\text{--}0.98$ ) and minor amounts of graphite can be used to constrain  $f_{\text{O}_2}$  during metamorphism. The ilmenite isopleths in  $T$ -log  $f_{\text{O}_2}$  space of Spencer & Lindsley (1981) suggest log  $f_{\text{O}_2} = -22$  to  $-17$  in the temperature interval 650–700°C. The maximum stability of graphite in  $T$ -log  $f_{\text{O}_2}$  space as depicted by Lamb & Valley (1985, fig. 3) indicates log  $f_{\text{O}_2}$  below  $-18$  to  $-17$  over the same range in  $T$  at pressures indicated by barometry.

The equilibrium



was used to calculate directly the  $f_{\text{O}_2}$  recorded by these phases in the paragneiss as suggested by the method of Bhattacharya & Sen (1986). The expression in Joules, bars, and degrees

K is:

$$\log f_{\text{O}_2} = \frac{\Delta H_{f,s}^0 - T\Delta S_{f,s}^0 + (P-1)\Delta V_{298}^0}{28.717T} + 0.869 \log(a_{\text{Fe}_2\text{O}_3}^{\text{ilm}}) - 0.579 \log(a_{\text{Fe}_3\text{Al}_2\text{Si}_3\text{O}_{12}}^{\text{grt}}).$$

where *s* designates solid phases. The internally consistent thermochemical data of Bhattacharya (1986) and the  $\text{Fe}_2\text{O}_3$  data of Robie *et al.* (1979) yield  $\Delta H_{f,s}^0 = -644\,442$  J,  $\Delta S_{f,s}^0 = 48.229$  J K<sup>-1</sup>, and  $\Delta V_{298}^0 = 5.133$  J b<sup>-1</sup>. The activity of  $\text{Fe}_2\text{O}_3$  in the ferrian ilmenite phase was calculated using the activity expression of Anderson & Lindsley (1981). Activity of  $\text{Fe}_3\text{Al}_2\text{Si}_3\text{O}_{12}$  in garnet was computed using the solution model of Ganguly & Saxena (1984). Garnet rim and matrix ilmenite compositions from sample FL-003 yield  $\log f_{\text{O}_2}$  values of  $-24$  at 3.5 kb and 730°C. This value is insensitive to the activity models used owing to the dominance of the first term in the  $\log f_{\text{O}_2}$  expression. Trial calculations using the heat capacity equations of Bhattacharya & Sen (1986) and Robie *et al.* (1979) show that correcting the thermochemical data for *T* greater than 298.15 K yields lower  $\log f_{\text{O}_2}$  values. The calculated  $f_{\text{O}_2}$  is in agreement with that indicated by the Spencer & Lindsley (1981) ilmenite isopleths in indicating  $\leq$ QFM conditions and is consistent with the occurrence of minor graphite.

The low  $a_{\text{H}_2\text{O}}$  together with low  $f_{\text{O}_2}$  recorded in the paragneisses precludes the existence of a  $\text{CO}_2$ -rich fluid phase during at least the latter stages of metamorphism (Lamb & Valley, 1985). Moreover, barring extensive influx of non-C–O–H volatile species (e.g.,  $\text{N}_2$ ); low  $a_{\text{H}_2\text{O}}$  and  $f_{\text{O}_2}$  requires fluid-absent conditions (Lamb & Valley, 1985). The paucity of channel-filling  $\text{H}_2\text{O}$  or  $\text{CO}_2$  in cordierite is consistent with these data in indicating that the observed paragenesis represents fluid-absent metamorphism.

#### PARTIAL MELTING

Portions of the paragneiss complex are migmatitic. Pressures and temperatures attending metamorphism coincide with estimated conditions for partial melting of pelitic rocks (Holdaway & Lee, 1977; Thompson, 1982; Newton, 1986). The migmatitic leucosomes are therefore considered to be *in situ* anatectic melts derived from the paragneisses. However, comparison of *T* and low  $a_{\text{H}_2\text{O}}$  recorded by mineral equilibria with experimental and calculated fluid-absent solidii for aluminosilicate melts (e.g., Bohlen *et al.*, 1983; Clemens & Vielzeuf, 1987) indicates that the preserved paragenesis in the pelitic samples examined was not in equilibrium with a melt phase. For example, solidii for aluminosilicate melt derived from the liquid solution model described by Burnham & Nekvasil (1986) and presented by Clemens & Vielzeuf (1987, fig. 2) show that temperatures  $\leq 750^\circ\text{C}$  (the range suggested by thermometry) require  $a_{\text{H}_2\text{O}} \geq 0.62$  for equilibrium with melt of the haplogranite system. At  $a_{\text{H}_2\text{O}}$  indicated by the paragneiss ( $\leq 0.26$ ) the solidus temperature is  $\geq 850^\circ\text{C}$ . Similar results are obtained by comparison with the experimental data for melting of phlogopite + quartz (Bohlen *et al.*, 1983, fig. 3).

Potential coexistence of hercynite and quartz could be interpreted to be indicative of  $T \geq 800^\circ\text{C}$  during an earlier portion of the metamorphic path (Bohlen *et al.*, 1983; Vielzeuf, 1983). However, the significant gahnite component in the spinels suggests stability with quartz could have occurred at lower temperatures (Bohlen & Dollase, 1983; Shulters & Bohlen, 1987). In any event, there is no evidence for coexisting hercynite and quartz in these rocks owing to the occurrence of spinel exclusively in cordierite. Hence, partial melting must have occurred prior to the development of the observed mineral assemblage and/or mineral

compositions when  $a_{\text{H}_2\text{O}}$  was significantly higher. Melting could have been initiated near the second sillimanite isograd at approximately 650 °C and pressures of 3–5 kb, producing small proportions of liquid while efficiently purging the remaining rock of  $\text{H}_2\text{O}$  (Thompson, 1982; Grant, 1985). A small degree of partial melting is indicated in the paragneisses by the persistence of abundant non-idiomorphic quartz (cf. Grant, 1985).

## DISCUSSION

Thermobarometry for the McCullough Range paragneisses indicates an apparent linear thermal gradient of  $\sim 50^\circ \text{ km}^{-1}$  in the Early Proterozoic crust of this region. This gradient corresponds to a surface heat flow  $q^0$  of  $\sim 122 \text{ mWm}^{-2}$  using reasonable upper-crustal thermal conductivities (Clark, 1966). Lambert (1983) calculated that equilibrium (steady state) surface heat flow for 30–40 km thick Early Proterozoic continental crust was roughly 84–88  $\text{mWm}^{-2}$ , with 108  $\text{mWm}^{-2}$  being a plausible maximum. The indicated isochronous geothermal gradients are 34, 36, and  $41^\circ \text{ km}^{-1}$  respectively. Based on these estimated steady state geotherms it can be concluded that the  $P$ – $T$  conditions recorded by the McCullough paragneisses required a transient (polychronic) thermal gradient. Three actualistic explanations for the anomalously high temperatures at shallow depth are: (1) displacement of peak temperatures induced by a perturbed geotherm to upper crustal levels by substantial uplift and erosion; (2) heating by anomalously shallow asthenosphere as a result of thinning of mantle lithosphere in an extensional tectonic regime; and (3) conductive heat supplied by invading mafic and felsic plutons. A brief discussion of the relative merits of these explanations as they apply to the McCullough complex follows.

Several workers have illustrated the importance of uplift and erosion in determining thermobarometric conditions in continental crust (Clark & Jager, 1969; Bickle *et al.*, 1975; England & Richardson, 1977; England & Thompson, 1984; Day, 1987). Appreciable denudation during metamorphism would result in displacement of peak temperatures experienced by rocks to higher crustal levels (lower pressures) (England & Thompson, 1984). If preceded by a period of isobaric heating and isostatic disequilibrium, denudation would promote recording of high temperatures by mineral equilibria at pressures significantly below that of the depth of origin (Thompson & England, 1984). An indication of the magnitude of this effect can be gleaned from the treatment of Bickle *et al.* (1975). These workers demonstrate the effects of perturbation of thermal gradients in the eastern Alps (Tauern Window) by uniform denudation. Their calculations suggest that an erosion rate of 1 mm/a operating for 30 Ma could produce an increase in surface heat flow on the order of  $51 \text{ mWm}^{-2}$  (Bickle *et al.*, 1975, fig. 8d). At face value this increase in  $q^0$  would have been sufficient to increase the temperature of Proterozoic crust at a depth of 15 km from the steady state value of  $\sim 500^\circ \text{C}$  (Lambert, 1983) to the metamorphic temperatures observed in the McCullough pelitic gneisses ( $\geq 700^\circ \text{C}$ ).

Wickham & Oxburgh (1985) and Sandiford & Powell (1986) have suggested that high-temperature low-pressure metamorphic terranes evolve in continental rift zones. The primary impetus for this correlation is the exceedingly high surface heat flow characteristic of present-day continental extensional environments. However, metamorphism in the McCullough Range and surrounding areas occurred during a major Early Proterozoic orogeny resulting in transcontinental addition of new continental crust at 1.74 to 1.68 Ga (Nelson & DePaolo, 1985). The influx of mantle-derived plutons was synchronous with regional compression (Karlstrom *et al.*, 1987). Although continental rifting is not tenable in this tectonic regime, localized extension may have occurred. Orrell *et al.* (1987) and Bender *et*

*al.* (1988) have suggested that Proterozoic rocks in the lower Colorado River region represent a rear-arc extensional tectonic setting. Evidence for a rear-arc setting includes synorogenic to late orogenic bimodal plutonism, which is anomalous with respect to calc-alkaline intermediate magmatism typical of the remainder of the orogen. If their model is correct, some lithospheric thinning and associated high heat flow may have contributed to low-pressure high-temperature metamorphism in the region.

Irrespective of the existence of an extensional environment, intrusion of voluminous granitic and mafic plutons beginning approximately 50 Ma prior to metamorphism, and continuing intermittently for about 90 Ma (Miller & Wooden, 1988), must have resulted in significant advective transport of heat to the upper crust. The granitic plutons are largely crustally derived. The isotopic data of Nelson & DePaolo (1985) and Bennett *et al.* (1988) indicate that the thermal perturbations which triggered crustal partial melting may have been caused by emplacement of mantle-derived plutons at greater depths. One-dimensional conductive heat flow modeling by Wells (1980) shows that with sufficient flux of magma ( $>0.8$  mm/a), under-plating and over-accretion of plutons may impart temperatures in excess of  $700^{\circ}\text{C}$  to the upper crust. The models also predict that these transient thermal gradients would persist for approximately 20 Ma following cessation of magmatism.

The apparent efficiency of pervasive plutonism in producing elevated polychronic thermal gradients implies that temporally coincident pluton emplacement was at least partially responsible for the Proterozoic granulite grade metamorphism. If the crustally derived plutons were caused by addition of mantle-derived melts at deeper levels, resultant thickening of the crust may have initiated uplift. The ensuing erosion would have also contributed to the low pressures recorded in the pelitic gneisses.

### CONCLUSIONS

Pelitic gneisses exposed in the McCullough Range of southern Nevada were metamorphosed during Early Proterozoic time at temperatures in excess of  $700^{\circ}\text{C}$  and at pressures of approximately 4 kb. The latter stages of metamorphism were characterized by subsolidus growth of cordierite at the expense of biotite and sillimanite. Lack of channel-filling fluid in cordierite, low  $a_{\text{H}_2\text{O}}$  (0.16), and low  $f_{\text{O}_2}$  ( $\leq \text{QFM}$ ) indicate fluid-absent conditions during this stage of metamorphism. Limited partial melting occurred prior to attainment of conditions recorded by geothermobarometry, resulting in desiccation of the pelitic gneiss.

Granulite grade metamorphism was in response to transient elevated thermal gradients produced in part by synchronous intrusion of mafic and felsic plutons during a period of significant continental accretion. Upper-crustal residence during high-temperature metamorphism may have been facilitated by attendant tectonic or erosional denudation of thickening crust.

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## APPENDIX

### *Analytical methods*

Approximately 200 mineral chemical analyses from four samples of garnet-cordierite gneiss were obtained using the ARL-SEMQ automated microprobe housed at the U.S. Geological Survey, Denver. Thirty four analyses from two additional samples were obtained with the MAC-5-SA3 microprobe at the California Institute of Technology.

The ARL-SEMQ microprobe utilizes six scanning wavelength-dispersive spectrometers simultaneously. An accelerating voltage of 15 kV was used for all analyses. Beam current was set to be equivalent to a sample current of 10 nA on brass. A spot size of approximately 10  $\mu\text{m}$  was used for alkali-poor minerals, and was increased to 20  $\mu\text{m}$  for feldspar and mica analyses. Counting time was 20 s for both peak and background measurements. Data reduction was performed utilizing the empirical method of Bence & Albee (1968). Averages of multiple analyses and single point analyses were used in this study. Multiple analyses (>4) of garnet and cordierite yielded sample standard deviations comparable to those derived from counting statistics. In most cases, major element precision indicated by multiple analyses was better than that suggested by counting statistics. For elements greater than 15 wt.% in abundance, the indicated relative precision at the 1 $\sigma$  confidence level is <1%. The relative precision at 1 $\sigma$  for elements ranging down to 5.0 wt.% is between 1 and 2%. Between 5.0 and 0.5 wt.%, relative precision at 1 $\sigma$  is 2–5%.

Operating conditions for the MAC-5-SA3 microprobe were comparable to those described above with the exception of counting times, which were determined by on-line computer to ensure less than 1% relative precision. No systematic error between the data collected from the two instruments was detected.