Multiple Metamorphic Events at Broken Hill, Australia. Evidence from Chloritoid-bearing Parageneses in the Nine-Mile Mine Region

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Pressure (P) and temperature (T) conditions for two distinct metamorphic events are quantified in the Nine-Mile Mine region of the central Broken Hill Block, New South Wales, Australia. Parageneses of the earlier event formed contemporaneously with D2 which produced a pervasive foliation. The event occurred at conditions around 650°C and 4–5 kbar, but may have been preceded by higher peak temperatures, possibly during an early M1 event. Thus, the event described here is called M2. This is supported by geochronological studies that report at least two high-grade partial melting events in the region. A later paragenesis, here called M3, overprinted statically the M2–D2 and D3 fabrics. PT conditions during M3 reached 480°C at 5 kbar. Possible PT paths between M2 and M3 are inferred from the interpretation of chloritoid-bearing parageneses using (1) a new calculated petrogenetic grid, (2) calculated compatibility diagrams and (3) thermodynamic pseudosections. In particular, two possible PT paths are discussed: (a) a direct path involving negligible pressure change during cooling and (b) cooling with minor decompression from M2 and much later compression and heating to M3. PT path (b) is interpreted from pseudosections. It is also considered more likely because: (1) the orientation and position of the regional isograds for M2 and M3 do not coincide across the Broken Hill Block and (2) D3 is interpreted as a substantial shortening event. Thus, a much larger P difference would be expected between M2 and M3 if the two events occurred on the same PT path. The interpretation of independent PT paths for M2 and M3 allows the possibility that the M3 event is much younger than the mid-Proterozoic high-grade events.

INTRODUCTION
The Broken Hill Block, New South Wales, Australia, hosts one of the largest lead–zinc–silver deposits in the world (e.g. Rahmdohr, 1950; Plimer, 1985). Because of its economic significance, a large amount of geological research has been conducted in the Proterozoic high-grade metamorphic rocks, the Willyama Supergroup, that hosts the deposit [for reviews see Hobbs et al. (1968, 1984), Vernon (1969) and Stevens (1986)]. Much of the work has been structural or stratigraphic, with less emphasis on the metamorphic evolution. Previous metamorphic studies have derived pressure–temperature (PT) conditions and attributed those to a single metamorphic event that changes in grade across the terrain (Binn, 1964; Phillips, 1980; Phillips & Wall, 1981) (Fig. 1a). Similarly, a single-event metamorphic PT path has been established for the Olary Block, that is adjacent to the Broken Hill Block to the west (Clarke et al., 1987). There, compression during cooling was inferred from interpreted superposition of lower-T, higher-P assemblages onto

KEY WORDS: Broken Hill; chloritoid equilibria; PT path; multiple events

Mineral abbreviations: bi, biotite; st, staurolite; ctd, chloritoid; sill, sillimanite; ky, kyanite; and, andalusite; as, aluminosilicate; g, garnet; chl, chlorite; mu, muscovite; cpx, clinopyroxene; opx, orthopyroxene; q, quartz. End-member abbreviations: mu, muscovite; cel, celadonite; mctd, Mg-chloritoid; fctd, Fe-chloritoid; mst, Mg-staurolite; fst, Fe-staurolite; phl, phlogopite; ann, annite; py, pyrope; alm, almandine; clin, clinohlore; ames, amesite.

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higher-$T$, lower-$P$ parageneses. This path was attributed to the ‘Olarian’ orogeny at ~1600 Ma. In the Broken Hill Block too, lower-grade metamorphic assemblages have long been recognized to overprint the high-grade peak assemblage (e.g. Edwards, 1958; Vernon, 1969; Vernon & Ransom, 1971; Corbett & Phillips, 1981), but only few studies (e.g. Hobbs et al., 1984; Stevens et al., 1990) have attempted either to correlate different parageneses with different thermal events, or to establish a $PT$ path from different parageneses. Such a study becomes increasingly important as more detailed evidence for a series of thermal events emerges from geo-
chronological studies. For example, thermal events accompanied by magmatism at ~1690 Ma, ~1640 Ma and ~1600 Ma have been resolved by SHRIMP U/Pb geo-
chronological studies (Page & Laing, 1992; Ehlers & Nutman, 1997; Nutman & Ehlers, 1997a,b), by conventional U/Pb dating (Gulson, 1984) and by Rb/Sr dating (Pidgeon, 1967). Younger thermal overprints at ~1200 Ma and ~500 Ma have been inferred from U/
Pb SHRIMP dating (Ehlers & Nutman, 1997) and Ar/Ar dating (Harrison & McDougall, 1981; Foster & K. Ehlers, unpublished data).

In this paper, $PT$ conditions for two overprinting metamorphic events, here called $M_3$ and $M_5$, are quan-
tified for the Nine-Mile Mine region, and possible paths between the two events are discussed. In particular, we present for the first time quantitative $PT$ estimates for the overprinting $M_3$ event that was first described by Edwards (1958) and later discussed by Vernon & Ransom (1971), Chenhall (1973) and Corbett & Phillips (1981). The ‘Nine-Mile Mine region’, as used here, encompasses the Living Desert Reserve and the regions around the Nine-Mile and Southern-Cross Mines in the region of grid reference 70±74N and 39±44E on the Broken Hill geological sheet (Brown, 1981) (Fig. 1b). It is ge-
ographically central to the Broken Hill Block and is located within 8 km of the ore body (Fig. 1a). Two methods are used to quantify the $PT$ history of the Nine-Mile Mine region: (1) average pressure–temperature calculations on mineral assemblages and (2) forward modelling of mineral assemblages through the $PT$ history by the use of cal-
culated compatibility diagrams. Peak conditions and different possible $PT$ paths are discussed in the light of independent regional field evidence and further con-
strained with petrogenetic pseudosections.
Geology of the Nine-Mile Mine region

According to Phillips (1980), the Nine-Mile Mine region is located within the ‘two-pyroxene zone’, the highest-grade zone of the Broken Hill Block. It is also located within the region of most intense ‘retrograde’ overprint (Chenhall, 1973; Brown, 1981) (Fig. 1b). According to Stevens et al. (1988), a regional metamorphic gradient of this retrograde overprint can be observed. This gradient is defined by the first appearances of late kyanite and staurolite (Fig. 1a). As part of this study, we have mapped out these isograds in some detail (Fig. 1b). The region comprises a variegated assembly of rock types including meta-pelitic schists, mafic gneisses, partial melt bodies and late crosscutting dykes (Stevens, 1978; Brown, 1981).

Volumetrically, the meta-pelitic units are the dominant rock types. They also host an abundance of Fe–Mg minerals that are useful for the quantitative derivation of a $PT$ path. Thus, the metamorphic descriptions of this study will focus on the meta-pelitic rocks.

In most of the region, evidence for two pervasive ductile deformation phases can easily be identified in the field. The intense penetrative fabric of the rocks is ascribed to an early deformation phase, in the literature known as $D_2$ (e.g. Marjoribanks et al., 1980). This deformation phase is associated with a strong $S_2$ foliation and $L_2$ lineation which are defined by minerals that grew during the earlier of the two metamorphic events ($M_2$) described below. A later deformation event folded this fabric into...
mesoscopic folds and microscopic crenulations. These mesoscopic folds characterize the regional geology and the responsible deformation phase is here called $D_3$. This nomenclature of the deformation events corresponds to the widely accepted nomenclature by Marjoribanks et al. (1980), and has recently been used by Gibson et al. (1996) and White et al. (1996), but it is noted that $D_3$ and $D_4$, as used here, correspond to $D_1$ and $D_2$ in the nomenclature of Hobbs et al. (1984). The region was subsequently cut by a series of retrograde schist zones (e.g., Vernon & Ransom, 1971), for example, the Stevens Creek Schist zone (Fig. 1b). This shear zone producing event is often referred to as $D_4$ in the literature. Further subdivision of the deformation history is unjustified in the light of the metamorphic focus of this study.

**METAMORPHIC GEOLOGY**

Two distinctly different parageneses can be observed in the field and in thin section. All minerals of the first assemblage are strictly aligned within the $S_3$ fabric (Fig. 3). Partial melt bodies are also aligned in this fabric. The second assemblage is a much coarser-grained porphyroblast assemblage. The porphyroblasts are essentially undeformed and overprint the $S_3$ fabric as well as the $D_3$ crenulations. Although we acknowledge the possibilities of mimetic growth and the problems associated with interpretation of reaction textures (e.g., Robinson, 1991; Vernon, 1996), we follow the classic textural interpretation and interpret the $S_3$ assemblage to have grown during $D_3$. Although this assemblage is the first equilibrated paragenesis that can be observed in thin section, we call it $M_2$ for two reasons: (1) at least two melt producing events are recognized within the study area (Nutman & Ehlers, 1997b) and the assemblage corresponds to the later of these two events; (2) this nomenclature emphasizes the synchronous nature of $D_3$ and $M_2$. The second assemblage appears largely static and is interpreted to have grown postdate to $D_3$. It is here referred to as $M_3$. The interpretation of the metamorphic events with the deformation phases does not preclude a model for the genetic relationship of the two events. Even though the two events are separated by $D_3$ (Fig. 2), they may form two periods of equilibration of one $PT$ path, or they may be unrelated in time and cause. The petrographic descriptions provided below will serve as a basis for the subsequent interpretation of this question. For petrographic descriptions of the rocks from the Nine-Mile Mine region we also refer to the detailed work of Browne (1922), Stillwell (1922) and Chenhall (1973).

![Fig. 2. Schematic sketch illustrating the nomenclature and timing relationships of deformation and metamorphic events used in this paper.](image)

Petrography

In all meta-pelitic rocks, the minerals biotite, muscovite and abundant quartz, as well as variable amounts of sillimanite, define a fine-grained $S_3$ fabric (matrix in Figs 3, 4a and d–g; Table 1). Garnet may be present. Minor amounts of tourmaline, apatite and magnetite are common. In low-strain zones, $M_3$ sillimanite and micas commonly occur as centimetre-sized aggregates of somewhat rectangular shape, possibly indicating the presence of pre-$M_3$ andalusite. The amount of sillimanite in the $S_2$ fabric is a poor indicator of the alumina content of the rock. In Al-poor rocks $M_3$ sillimanite may be preserved within $M_3$ quartz grains or as inclusion trails in the poikiloblastic assemblage of the second assemblage (Fig. 4f). Al-rich rocks, on the other hand, may contain abundant aluminous $M_1$ minerals, but little free sillimanite.

The second ($M_3$) assemblage is characterized by static overgrowth of garnet, chloritoid, staurolite and chlorite porphyroblasts over the $M_3$ assemblage. These porphyroblasts constitute 5–20 vol. % of the rocks and individual porphyroblasts generally range in size between 1 mm and 1 cm, although garnet, staurolite and chloritoid grains of up to 5 cm were found. In contrast to the uniform $M_3$ assemblages, the $M_3$ porphyroblast assemblage is dependent on the alumina content of the rock (Table 1). In very aluminous rocks (‘Bulk 1’, see below) the poikiloblastic assemblage is staurolite–chloritoid (Figs 3a and 4a). In rocks with intermediate alumina content (‘Bulk 2’) it is either chloritoid–chlorite or garnet–staurolite (Fig. 4c) or garnet–chlorite–chloritoid, and in rocks with low alumina content (‘Bulk 3’) it is garnet–chlorite (Figs. 3b and 4b). Most porphyroblasts are idiomorphic and completely overgrow the mica fabric, but preserve inclusion trails of oxides and sillimanite, which still record the $S_3$ fabric. The static nature of this porphyroblast growth is evidenced by pseudomorphs of staurolite and garnet after biotite (Fig. 4d and e) and staurolite after garnet. Apart from these porphyroblasts, $M_3$ is characterized by pervasive muscovite growth in some rocks (Fig. 4b and g–i).
Characteristics of bulk compositions and petrography

Qualitative parageneses and quantitative mineral compositions show that all meta-pelitic rocks are extremely Fe rich, usually with $X_{Fe}>0.8$ [$X_{Fe} = Fe/(Fe+Mg+Mn)$], with some rocks being almost Mg free. However, the rocks vary greatly in their alumina content. Depending on the alumina content, three different bulk compositions may be outlined (Table 1) which have characteristic $M_3$ assemblages. ‘Bulk 1’ are rocks with $0.5<X_{Al}<0.7$ [where $X_{Al} = Al/(Al+Mg+Fe)$, molar], which places them between chloritoid and staurolite in the AFM projection. ‘Bulk 2’ are rocks with an alumina content of $0.25<X_{Al}(<0.5$, which places them between garnet and chloritoid, and ‘Bulk 3’ are rocks that are less aluminous than garnet ($X_{Al}<0.25$) (Fig. 5a).

Very aluminous rocks consist largely of sillimanite with only small amounts of Fe–Mg minerals, and very alumina-poor bulk compositions contain only biotite–muscovite, occasionally with abundant magnetite and ilmenite. However, these extreme bulk compositions are of high thermodynamic variance and therefore have similar $M_2$ and $M_3$ parageneses. Thus, they are not useful for the derivation of a $PT$ path and will not be discussed further. Similarly, meta-pelitic rocks of extremely unusual bulk compositions containing an abundance of rare minerals, for example, galinite, have been found around some of the mines, but

![Sketches of textural relationships in two characteristic bulk compositions: (a) Bulk 1 with $X_{Fe}^{Bulk1}>X_{Fe}^{Bulk1}$, and (b) Bulk 3 with $X_{Fe}^{Bulk3}>X_{Fe}^{Bulk3}$.](image)
Table 1: Characteristic meta-pelitic assemblages from the Nine-Mile Mine region and interpreted sequence of reaction

<table>
<thead>
<tr>
<th>Observed complete assemblage</th>
<th>Observed in $S_2$ fabric</th>
<th>Observed as $M_2$ overgrowths</th>
<th>Inferred equilib. $M_2$ assemblage</th>
<th>Inferred equilib. $M_3$ assemblage</th>
<th>Thin section examples</th>
<th>Best projection</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Bulk composition 1</strong></td>
<td>ctd-st –bi-mu–sill (g-chl)</td>
<td>ctd-st-(mu)</td>
<td>g-bi-mu-sill</td>
<td>g-st-ctd-mu-ky</td>
<td>19A, 19B</td>
<td>FMK + sill</td>
</tr>
<tr>
<td><strong>Bulk composition 2</strong></td>
<td>ctd-chl-bi-mu-sill</td>
<td>bi-mu-sill</td>
<td>g-ctd-chl-st-(mu)</td>
<td>g-bi-mu-sill</td>
<td>19A</td>
<td>FMK + g</td>
</tr>
<tr>
<td></td>
<td>g-st-bi-mu-(sill)</td>
<td>bi-mu-(sill)</td>
<td>g-st-(chl)</td>
<td>g-bi-mu-sill</td>
<td>18B, 19C</td>
<td>AMK + g</td>
</tr>
<tr>
<td></td>
<td>st-bi-mu-(sill)</td>
<td>bi-mu-(sill)</td>
<td>st</td>
<td>g-bi or g-bi-mu</td>
<td>18A</td>
<td>AMK + g</td>
</tr>
<tr>
<td><strong>Bulk composition 3</strong></td>
<td>g-chl-bi-mu-sill</td>
<td>bi-mu-sill</td>
<td>g-chl-mu</td>
<td>(g)-bi-mu-sill</td>
<td>19D</td>
<td>AFM + mu</td>
</tr>
<tr>
<td></td>
<td>g-chl-bi-mu</td>
<td>bi-mu</td>
<td>g-chl-mu</td>
<td>bi-mu</td>
<td>19E</td>
<td>AFM + mu</td>
</tr>
<tr>
<td><strong>Other common bulk compositions</strong></td>
<td>sill-(bi-mu)</td>
<td>sill-(bi-mu)</td>
<td>sill-(bi-mu)</td>
<td>sill-(bi-mu)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>bi-mu-(sill)-mt</td>
<td>bi-mu-(sill)-mt</td>
<td>bi-mu-(sill)-mt</td>
<td>bi-mu-(sill)-mt</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Phases that occur in only minor or accessory amounts are in parentheses; phases that are volumetrically abundant are given in italics. The first three columns refer to the actually observed minerals. The last two columns are the equilibrium assemblages of $M_2$ and $M_3$ interpreted from the observations.

are not discussed here. The discussion of individual minerals given below is presented in the order of the interpreted time of their stability.

Biotite constitutes ~10–30% of the rock volume. Biotite grains are aligned in the $S_2$ fabric but some large crystals may cut across it. Biotite is interpreted as part of the $M_2$ assemblage. Only in Al-poor bulk compositions is biotite in textural equilibrium during the static growth of the $M_2$ assemblage. In other bulk compositions, $M_2$ biotite is severely corroded by $M_1$ minerals, in particular muscovite (Fig. 4h). For example, biotite is practically absent in $M_1$ chloritoid-bearing rocks. Crosscutting biotite crystals are interpreted to have recrystallized at the end of $M_2$, before being destabilized by the $M_3$ conditions.

Sillimanite occurs in fibrolitic and in prismatic habit and is the $S_2$ fabric defining mineral. It is interpreted to be stable during $M_3$. Abundant sillimanite is also present during $M_2$, although $M_2$ is within the kyanite stability field (see below). $M_3$ sillimanite must therefore be preserved metastably.

Garnet occurs as strongly corroded and deformed crystals within the $S_2$ fabric and as idiomorphic overgrowths over the $S_2$ fabric in spectacular pseudomorphs, for example, after biotite. Garnet pseudomorphs after biotite preserve not only the $S_2$ fabric (Fig. 4c), but also the geometry of individual biotite crystals in the form of rutile needles (Fig. 4d). $M_2$ garnets are microscopically easily recognized by strong corrosion features and highly deformed inclusion trails (Fig. 3b). Garnet is therefore interpreted as part of the $M_2$ and the $M_3$ assemblages. In the field, $M_2$ garnets are up to several centimetres long and are easily recognized by their strongly elongated shape in $S_2$.

Muscovite occurs as small crystals aligned in the $S_2$ fabric, as random overgrowths over the $S_2$ fabric and as large porphyroblasts that grew during $M_3$. Some $M_3$ muscovite porphyroblasts corrode $M_1$ staurolite, garnet, chlorite and chloritoid. Thus, muscovite blastesis is interpreted as the last event of $M_3$, but muscovite that grows in the $S_2$ fabric is interpreted as part of the $M_3$ assemblage. Random muscovite growth at the expense of biotite is widespread in most thin sections and has gone to various stages of completion (Figs 3 and 4i).

Chloritoid occurs as millimetre- and centimetre-sized idiomorphic $M_3$ porphyroblasts in bulk compositions 1 and 2. Usually chloritoid contains fine and coarse sillimanite inclusions preserving $S_2$ (Fig. 4f). In bulk composition 1, it occurs in textural equilibrium with staurolite. Occasionally, it grows around staurolite, suggesting that it grew cogenetic with, or slightly after staurolite (Fig. 4a). In bulk composition 2, chloritoid occurs in textural equilibrium with millimetre-sized chlorite crystals.

Staurolite occurs only as idiomorphic porphyroblasts. Crystals are up to several centimetres in size. Some staurolites appear cubic in shape and are interpreted as...
Fig. 4. Photomicrographs of characteristic porphyroblast-matrix relations and reaction textures in meta-pelitic assemblages. The three characteristic $M_3$ porphyroblast assemblages are shown in: (a), st-ctd (Bulk 1); (b), g-chl (Bulk 2); (c), st-g (Bulk 3). Most photographs show the strong $M_2$ bi-mu-(sill) fabric in the matrix outside the porphyroblasts and inclusion trails parallel to this fabric inside the porphyroblasts. Static overgrowth of garnet, staurolite and chloritoid porphyroblasts over $M_2$ biotite-sillimanite is spectacularly shown in (d), (e) and (f). In (d) interlocking biotite geometry can still be identified by rutile needles inside the garnets. (g) shows $M_3$ staurolite being corroded by late-stage muscovite. The shift of the modal proportions of biotite and muscovite in favour of muscovite during $M_3$ can be seen in the background of (b), (c), (e) and (g) but is particularly shown in (h), where an $M_3$ muscovite porphyroblast replaces biotite, and in (i), which shows the typical random $M_3$ muscovite growth. Width of photographs (a)–(f) and (h) is ~1 mm, and ~0.5 mm for (g) and (i). Micrographs are from the following thin sections: (a), (c), (i): BH19A; (f): BH19B; (e), (h): BH19C; (d): BH19D; (b): BH19E.
Fig. 5. (a) Division of the AFM projection in three characteristic bulk compositions. (b) Mineral compositions of the characteristic meta-pelitic parageneses from the Nine-Mile Mine region in the AFM projection. The homogeneity of the compositions is indicated by the size of the compositional fields. The staurolite field encompasses all 35 staurolite analyses from cores and rims of five thin sections. Chlorite field shows 10 analyses from two thin sections, garnet field shows 43 analyses from five thin sections, chloritoid is for a total of 10 analyses from two thin sections. Circles define the qualitative position of the bulk composition of the various thin sections. Because of the projection from different muscovite compositions, the $X$Fe spread of biotite appears larger than it is.

pseudomorphs after garnet (Fig. 4g). They often preserve inclusion trails mimicking the $S_2$ fabric (Fig. 4a and c) or cleavages of former biotite (Fig. 4e). In the latter case, the inclusion trails are made up of rutile, which incorporates Ti present in biotite but not present in staurolite. Staurolite may be partly corroded by muscovite (Fig. 4g). All staurolite is interpreted as part of the $M_3$ assemblage.

Chlorite porphyroblasts are idiomorphic and they grow up to several millimetres in size. Like staurolite, chlorite is often overgrown by last-stage muscovite growth (Fig. 4b). Chlorite is also interpreted as part of the $M_4$ assemblage.

Mineral chemistry
About 100 thin sections of meta-pelites were studied optically. All of them are practically identical in their three characteristic assemblages and reaction textures. Therefore only 10 thin sections representing the three characteristic bulk compositions were selected for microprobe analyses. About 400 high-quality analyses were produced to cover all minerals in each assemblage. All analysed grains are of remarkably similar composition to the M3 assemblage.

Chloritoid porphyroblasts are idiomorphic and they grow up to several millimetres in size. Like staurolite, chlorite is often overgrown by last-stage muscovite growth (Fig. 4g). All staurolite is interpreted as part of the $M_3$ assemblage.

Analytical technique
Mineral analyses were performed on a four-spectrometer ARL-SEM electron microprobe at the Department of Earth Science at Monash University with an acceleration potential of 15 kV and a sample current of 15 nA. Electron beam spot size was 1 μm for garnet, staurolite,
chloritoid and chlorite analyses, and 10 μm for micas and occasionally for chlorite. Only wavelength-dispersive analysis were obtained and ZAF corrections applied. Minerals were checked for homogeneity with back-scattered images. Silicates were used as standards. Representative mineral analyses are shown in Table 2.

Mineral chemistry
The 56 analyses obtained of biotite grains vary in the range 0.53<XFe<0.71, with most analyses being higher than XFe>0.60 (Fig. 5b). Biotite often contains between 1 and 2 wt % TiO2 and ~0.5 wt % F. There appears to be no systematic variation of the biotite composition with bulk composition or assemblage.

Garnet is also completely unzoned and largely homogeneous in composition, with all 43 analyses obtained falling in a narrow compositional range of 0.93<XFe<0.95. Only in extremely Fe-rich garnet-clast assemblages, garnet may reach XFe=0.959. In all thin sections garnet is more Fe rich than staurolite. Manganese contents around 3–5 wt % and calcium contents between 1 and 2 wt % are common.

The 58 muscovite analyses obtained vary in the range 0.31<XFe<0.85 but 90% of all analyses are around 0.71<XFe<0.75. Optically different muscovite generations cannot be distinguished compositionally. Total FeO+MgO content rarely exceeds 3 wt %, and ~1 wt % Na2O is common. Most Fe-poor analyses of muscovites are from rocks containing staurolite as the only M1 phase.

Twelve high-quality analyses were obtained of chloritoid. Totals of these analyses vary between 91 and 93 wt %, indicating 7–9 wt % volatiles are present. Compositions are homogeneous, with XFe in the range 0.89<XFe<0.91 in assemblages where chloritoid occurs together with staurolite and garnet and 0.86<XFe<0.87 in assemblages where chloritoid occurs together with staurolite only (Fig. 5).

Staurolite is unzoned, has practically identical compositional variations in all bulk compositions, and all 35 analyses obtained from five thin sections fall in a narrow range of 0.87<XFe<0.91, with most analyses being in the range 0.89<XFe<0.90. MnO and TiO2 content are generally below 0.5 wt %, and ZnO content reaches 1 wt % in some analyses.

Chlorite porphyroblasts in bulk composition 3 are in the range 0.59<XFe<0.63 in 10 analyses. In chloritoid–chlorite assemblages of bulk composition 2, chlorite is somewhat more magnesian, with most chlorites being around XFe=0.56. This bulk composition dependence is consistent with calculated mineral equilibria which predict that chlorite in chloritoid–chlorite–garnet assemblages is slightly more magnesian than that in equilibrium with garnet and biotite (see Fig. 7a, below). Totals are generally between 86 and 87 wt %. Manganese, titanium and fluorine contents are negligible and fluorine is at the limit of detectability.

THE APPROPRIATE PETROGENETIC GRID
Table 2 shows that the aluminous minerals are made up to >95 wt % of the components K2O, FeO, MgO, Al2O3, SiO2, and H2O. Garnet contains up to ~4 wt % MnO. This is the largest amount of another component present in any of the minerals. Staurolite contains up to 1% ZnO. We begin therefore our considerations using a calculated petrogenetic grid in the idealized pelitic system KFMASH and consider the influence of MnO and ZnO on the stability relationships in the discussion. A grid was calculated for the phases chloritoid, staurolite, garnet, chlorite, biotite, muscovite, quartz, the aluminosilicate polymorphs and H2O using the computer software THERMOCALC (Powell & Holland, 1988) and an updated version of the thermodynamic dataset of Holland & Powell (1990) (T. J. B. Holland & R. Powell, personal communication, 1995) (Fig. 6). In this grid the FeMg±1 and the AlSi3(Fe,Mg)2 substitutions in muscovite, biotite and chlorite, and the FeMg±1 substitution for chloritoid, garnet and staurolite are considered. The formulation of compositional parameters, end-member activities and method of calculation are the same as those of Xu et al. (1994) and Stüwe & Powell (1995). Phillips (1980) suggested water activities below unity during peak metamorphism of the region of consideration. Thus, most equilibria at high T were calculated at a series of water activities.

In the PT range shown, the grid contains two stable and one metastable invariant point. At the chloritoid-out invariant point, [ctd], most reactions relevant to ‘normal’ meta-pelitic rocks meet, including the staurolite isograd, the staurolite–chlorite breakdown reactions and the staurolite–biotite breakdown reaction. The other invariant point, [bi,mu], is degenerate with respect to K2O. Two important univariant reactions involving chloritoid extend from this invariant point to lower T. These are the chloritoid–kyanite [g] and the chloritoid [as] breakdown reactions (with quartz in excess). The final chloritoid breakdown occurs at temperatures between 550 and 600°C in a pressure range between 5 and 12 kbar. The calculated grid remains questionable only with respect to the topology of the staurolite–biotite [ch] breakdown reaction. In most natural rocks, including the ones discussed here, staurolite has a lower XFe than garnet, indicating a topology of st+mu=g+as+bi, rather than st+bi=g+as+mu, as shown in Fig. 6. However, because the [bi,mu] reaction is degenerate with respect to K2O, the topology of the grid remains unaffected by the stoichiometry of this reaction.
Table 2: Representative microprobe analyses from characteristic minerals of the Nine-Mile Mine region

<table>
<thead>
<tr>
<th>Sample:</th>
<th>19A 19A 19A 19A 19A 19B 19C 19C 19C 19C 19C 19E 19E 19E 19E</th>
</tr>
</thead>
<tbody>
<tr>
<td>Analysis no.:</td>
<td>c2p2 c2p4 c2p8 c2p10 c2p7 c1p22 c3p10 c1p12 c3p1 c1p16 c3p6 c1p10 c19Ap6 c19Ap4 c19Ap10 c22Ap3</td>
</tr>
<tr>
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</tr>
<tr>
<td>TiO2</td>
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</tr>
<tr>
<td>Al2O3</td>
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<tr>
<td>MgO</td>
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<tr>
<td>CaO</td>
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</tr>
<tr>
<td>MnO</td>
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</tr>
<tr>
<td>FeO*</td>
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<tr>
<td>ZnO</td>
<td>0.48</td>
</tr>
<tr>
<td>Na2O</td>
<td>—</td>
</tr>
<tr>
<td>K2O</td>
<td>—</td>
</tr>
<tr>
<td>F</td>
<td>—</td>
</tr>
<tr>
<td>Total Oxygeny</td>
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<tr>
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<tr>
<td>Mg</td>
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<tr>
<td>Ca</td>
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<tr>
<td>Mn</td>
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<tr>
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</tr>
<tr>
<td>Zn</td>
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<tr>
<td>Na</td>
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<tr>
<td>K</td>
<td>0.000</td>
</tr>
<tr>
<td>F</td>
<td>0.000</td>
</tr>
</tbody>
</table>

Cations are calculated for all elements analysed. The analyses shown are those used for the PT calculation in Table 3 plus some additional representative analyses from a total of more than 400 analyses obtained.

*FeO measured as total Fe.
Fig. 6. Petrogenetic grid in the system KFMASH for the phases staurolite, garnet, chlorite, chloritoid, biotite, muscovite, quartz, H₂O and the aluminosilicate polymorphs as calculated with the software THERMOCALC. This grid contains a total of nine invariant points of which only [ctd] and the degenerate [bi,mu] are stable in the PT range shown. The only metastable invariant point in the PT range shown is [chl] invariant, which is shown for reference purposes. AFM compatibility diagrams are projected from muscovite, quartz and H₂O. Because of the projection from muscovite, not all are relevant to the K₂O-degenerate reactions of the grid. Black arrows indicate displacement of equilibria if MnO is considered, white arrows if ZnO is considered. Black regions show probable PT conditions as estimated from pseudosection information of Fig. 8. Grey shaded region shows the topologically possible area through which a PT path between M₂ and M₃ must cross. Metastable reactions shown by dashed lines, stable reactions by full lines. Dashed reactions parallel to the muscovite breakdown are the muscovite breakdown reaction at $\alpha$(H₂O) = 0.8 and 0.7. The metastable reaction ctd + sill = st + chl within the kyanite stability field is metastably seen by the rocks discussed here. Black arrows indicate displacement of equilibria if MnO is considered, white arrows if ZnO is considered.

If MnO is added to the KFMASH system, this will be largely incorporated into garnet and secondly into chloritoid (Table 2), increasing the stability range of these two phases. Thus, adding the effects of MnO to the grid of Fig. 6 will cause shifts of the invariant and univariant equilibria in the directions indicated by the small black arrows (Fig. 6). This shift will be, in part, counteracted by the effects of ZnO, which stabilizes the existence of staurolite in an opposite part of PT space (white arrows in Fig. 6). The topology of equilibria involving chloritoid has been subject to some controversy (Wang & Spear, 1991; Droop & Harte, 1995) and is a function of the topology around the [as] invariant point (below the PT range shown in Fig. 6). The topology given by THERMOCALC is consistent with that suggested by Harte & Hudson (1979). Accordingly, the effects of manganese on the calculated grid of Fig. 6 will correspond to those suggested by Droop & Harte (1995).

To illustrate the evolution of the three bulk compositions discussed above, quantitative compatibility...
diagrams were calculated (Fig. 7a–c). Compatibility diagrams for the three bulk compositions are projected from different phases to obtain maximum information for each paragenesis. For example, reaction textures involving modal changes of muscovite cannot be interpreted in projections from muscovite (Fig. 7a). Thus, projection from garnet onto the MgO–Al₂O₃–K₂O plane (Fig. 7b) was chosen to interpret phase equilibria in garnet-bearing rocks, and projection from aluminosilicate onto the FeO–MgO–K₂O plane (Fig. 7c) was chosen for rocks containing Al₂SiO₅ during M₂ and M₃.

For all projections, ternary compatibility diagrams were calculated between each of the univariant reactions shown in Fig. 6. In particular, compatibility diagrams were calculated at 525°C and 6 kbar; 540°C and 6 kbar; 565°C and 5.5 kbar; 585°C and 5.6 kbar; 600°C and 5 kbar; 608°C and 4.5 kbar and finally at 625°C and 4 kbar. However, it is emphasized that the compatibility diagrams are topologically equivalent for any chosen conditions within the six characteristic reaction spaces. Thus, the compatibility diagrams are representative for interpreting any PT path within the light-shaded area shown in Fig. 6.

These compatibility diagrams (and others) were then used to construct quantitative petrogenetic pseudosections for bulk compositions containing chloritoid. Two representative bulk compositions were chosen. ‘Bulk A’ is somewhat more aluminous than chloritoid and lies therefore between chloritoid and staurolite in AFM. ‘Bulk B’ is somewhat less aluminous than chloritoid and lies between garnet and chloritoid (Fig. 7a). Pseudosections are shown in Fig. 8a and b. These pseudosections are useful for the interpretation of bulk composition 2 and possibly bulk composition 1. At temperatures above the chloritoid breakdown, the two pseudosections are topologically almost identical. Below the chloritoid breakdown, different di- and trivariant chloritoid-bearing equilibria appear. It should be noted that all equilibria below the chloritoid breakdown reaction are degenerate with respect to muscovite and biotite. Thus, although muscovite does appear in the fields of higher variance, it does not take part in the reaction and volumetric changes are therefore not involved.

**METAMORPHIC CONDITIONS FOR M₂ AND M₃**

Minimum metamorphic conditions for M₂ are given by the stable coexistence of biotite–muscovite–sillimanite with or without garnet. In the presence of quartz and H₂O, this paragenesis defines peak conditions above the
reaction for both possible topologies of this reaction (Fig. 6). It also must lie within the sillimanite stability field of the aluminosilicate polymorphs. An upper limit may be given by the muscovite-quartz breakdown reaction, which has a positive slope in PT and lies at ~3 kbar at 650°C and ~6 kbar at 750°C (Fig. 6). This constrains the peak metamorphic conditions to lie between 600 and 700°C, and 3 and 6 kbar. Higher pressures are, in theory, possible. However, for the range of bulk compositions of the Nine-Mile Mine region, the boundary between the

---

Fig. 7. Calculated compatibility diagrams at five pressures and temperatures that are shown as filled circles in Fig. 6. (a) is projected from muscovite, quartz and H₂O onto the plane Al₂O₃-FeO-MgO and is the classic AFM projection; (b) is projected from garnet, quartz and H₂O onto the plane MgO-Al₂O₃-K₂O; and (c) is projected from aluminosilicate, quartz and H₂O onto the plane FeO-MgO-K₂O. Filled circles in (a) show the bulk compositions chosen for the pseudosections of (a) and (b). Labelled circles are bulk compositions of different thin sections and are only shown for those PT steps that are recorded by the parageneses.
Fig. 8. Petrogenic pseudosections of the grid shown in Fig. 6 for two characteristic bulk compositions involving chloritoid-bearing parageneses. Dark shaded fields are trivariant fields, white areas are divariant fields. Seen sections of univariant reactions are shown as thick lines. (a) is for an Fe-rich bulk composition with an alumina content between chloritoid and staurolite. The chosen bulk composition was defined by the compositional point at: mctd:fctd:fst:mu = 1:9:10:10. The two light shaded fields are the metastable positions of the ctd±sill±st±mu and ctd±sill±chl±mu divariant fields which are preserved metastably in the Nine-Mile Mine rocks. (b) is for a bulk composition within the AFM divariant field chloritoid, garnet and chlorite. The chosen bulk composition was defined by the compositional point mctd:fctd:py:alm:mu = 1:9:1:9:10.

divariant field g±bi±mu±sill and the trivariant field bi±mu±sill is never higher than 4–5 kbar (in fact, it is at 2.8 kbar for both bulk compositions shown in Fig. 8). Thus, the absence of garnet in many rocks indicates that $M_2$ peak conditions may be constrained to lie in a narrow range above 2.7 kbar, below 5 kbar and between 620 and 680°C (Fig. 6, black $M_2$ triangle). For water activities less than unity, the muscovite breakdown reaction slides to lower $T$ by about 15°C for every $a(H_2O)$ interval of 0.1. Phillips (1980) suggested a value of $a(H_2O) = 0.7$ for the region discussed here. This value constrains $M_2$ conditions to almost one $PT$ point at ~630 ± 10°C and
just below 4 kbar (Fig. 6). A possible pre-history of higher peak metamorphic conditions is examined in the discussion. Mineral compositions of $M_2$ have all re-equilibrated during $M_1$ and attempts to reproduce the interpreted $M_3$ conditions with quantitative thermobarochemistry on the $M_1$ assemblage failed.

Maximum metamorphic conditions for $M_3$ are given by both the interpreted stable coexistence of chloritoid–quartz during $M_3$ (placing $M_3$ conditions below the $[g]$ reaction in Fig. 6) and the interpreted coexistence of chloritoid–muscovite $+$ staurolite $+$ sillimanite (placing $M_1$ conditions into the respective trivariant field in Fig. 8a). The equilibrium chloritoid–sillimanite is metastable with respect to kyanite. However, textural evidence suggests that this equilibrium is preserved metastably. This metastable reaction has a $PT$ position $\sim 15^\circ C$ below the $ky + ctd = st + chl + q$ reaction (Fig. 6). Thus, $M_3$ can be constrained to lie below 500°C at 4 kbar or below $\sim 550^\circ C$ at 10 kbar. From the grid of Fig. 6 and the pseudosections of Fig. 8 higher pressures are theoretically possible, but we will show below that they are improbable.

The variation of mineral composition with paragenesis is consistent with the interpreted formation conditions. Chloritoid is slightly more Fe rich when occurring together with staurolite and garnet than it is when occurring with staurolite only (see mineral chemistry section). This is consistent with the compositions of calculated $M_3$ equilibria for the KFMASH divariant assemblage $ctd$–$g$–$st$–$mu$–$q$–$H_2O$ and the KFMASH trivariant assemblage $ctd$–$st$–$mu$–$q$–$H_2O$, at temperatures below $550^\circ C$ (Fig. 7a, bottom right compatibility diagram). Correspondingly, garnet is almost pure almandine, when in equilibrium with staurolite. This is consistent with the $M_3$ assemblage being KFMASH trivariant $(g$–$st$–$mu$–$q$–$H_2O)$ at temperatures below $550^\circ C$. Finally, chlorite is slightly more magnesian in equilibrium paragenesis with chloritoid and garnet than in paragenesis with biotite and garnet. This too is consistent with the calculated phase equilibria in the respective divariant fields in Fig. 7a at temperatures below $540^\circ C$.

Average temperature calculations on $M_3$ assemblages with and without chloritoid (Table 3) give temperatures between 438 and $530^\circ C$ ($\pm 50^\circ C$) for a chosen $P$ range between 3 and 8 kbar. However, the fit parameter, $f$, is lowest for pressures around 3–6 kbar, indicating these pressures as probable. Average $P$ and $T$ calculations on $M_3$ assemblages give similar temperatures, but no reliable constraints on $M_3$ pressures.

### THE $PT$ PATH BETWEEN $M_2$ AND $M_3$

The rocks of the Nine-Mile Mine region give a complete record of the $PT$ evolution between $M_2$ and $M_3$. We discuss it individually for the three characteristic bulk compositions.

Bulk composition 1 is the most aluminous bulk composition and sillimanite is present at all stages of its evolution (Table 1). Therefore a ternary projection from aluminosilicate is used to discuss its path from $M_2$ to $M_3$.

Bulk composition 1 plots in the trivariant biotite–muscovite or the divariant biotite–muscovite–garnet fields (plus sillimanite $+$ quartz $+$ $H_2O$) at $M_2$ conditions, for example, at 625°C and 4 kbar (Fig. 7c). During cooling, the first encountered reaction is the breakdown of garnet–muscovite to staurolite–biotite. However, little change of the modal proportions of biotite and muscovite is expected and staurolite grows mainly at the expense of chloritoid. The equilibrium biotite–sillimanite is metastable just below 4 kbar (Fig. 6). A possible pre-history of discussions is examined in the discussion. Mineral compositions of $M_2$ have all re-equilibrated during $M_1$ and attempts to reproduce the interpreted $M_3$ conditions with quantitative thermobarochemistry on the $M_1$ assemblage failed.

Maximum metamorphic conditions for $M_3$ are given by both the interpreted stable coexistence of chloritoid–quartz during $M_3$ (placing $M_3$ conditions below the $[g]$ reaction in Fig. 6) and the interpreted coexistence of chloritoid–muscovite $+$ staurolite $+$ sillimanite (placing $M_1$ conditions into the respective trivariant field in Fig. 8a). The equilibrium chloritoid–sillimanite is metastable with respect to kyanite. However, textural evidence suggests that this equilibrium is preserved metastably. This metastable reaction has a $PT$ position $\sim 15^\circ C$ below the $ky + ctd = st + chl + q$ reaction (Fig. 6). Thus, $M_3$ can be constrained to lie below 500°C at 4 kbar or below $\sim 550^\circ C$ at 10 kbar. From the grid of Fig. 6 and the pseudosections of Fig. 8 higher pressures are theoretically possible, but we will show below that they are improbable.

The variation of mineral composition with paragenesis is consistent with the interpreted formation conditions. Chloritoid is slightly more Fe rich when occurring together with staurolite and garnet than it is when occurring with staurolite only (see mineral chemistry section). This is consistent with the compositions of calculated $M_3$ equilibria for the KFMASH divariant assemblage $ctd$–$g$–$st$–$mu$–$q$–$H_2O$ and the KFMASH trivariant assemblage $ctd$–$st$–$mu$–$q$–$H_2O$, at temperatures below $550^\circ C$ (Fig. 7a, bottom right compatibility diagram). Correspondingly, garnet is almost pure almandine, when in equilibrium with staurolite. This is consistent with the $M_3$ assemblage being KFMASH trivariant $(g$–$st$–$mu$–$q$–$H_2O)$ at temperatures below $550^\circ C$. Finally, chlorite is slightly more magnesian in equilibrium paragenesis with chloritoid and garnet than in paragenesis with biotite and garnet. This too is consistent with the calculated phase equilibria in the respective divariant fields in Fig. 7a at temperatures below $540^\circ C$.

Average temperature calculations on $M_3$ assemblages with and without chloritoid (Table 3) give temperatures between 438 and $530^\circ C$ ($\pm 50^\circ C$) for a chosen $P$ range between 3 and 8 kbar. However, the fit parameter, $f$, is lowest for pressures around 3–6 kbar, indicating these pressures as probable. Average $P$ and $T$ calculations on $M_3$ assemblages give similar temperatures, but no reliable constraints on $M_3$ pressures.

Bulk composition 2 is less aluminous and sillimanite is not always present. However, garnet is present during $M_3$ and $M_2$ in most rocks of ‘Bulk 2’ and thus a projection from garnet onto the MgO–K$_2$O–Al$_2$O$_3$ plane was used (Fig. 7b). $M_2$ assemblages that consist of biotite–sillimanite–garnet and biotite–sillimanite–muscovite–garnet are illustrated in Fig. 7b (625°C and 4 kbar).

Bulk composition 3 is the least aluminous of the three and is characterized by the $M_3$ overprinting assemblage garnet–chlorite. In most parageneses of ‘Bulk 3’ muscovite and garnet are always present and therefore projections from muscovite onto Al$_2$O$_3$–Fe$_2$O$_3$–MgO (AFM) and from garnet onto MgO–K$_2$O–Al$_2$O$_3$ (MKA) are used. Figure 7a shows that the trivariant $M_3$ assemblage garnet–biotite (plus muscovite, quartz and $H_2O$) or the divariant garnet–biotite–sillimanite assemblage will form garnet–chlorite–biotite or chlorite–biotite assemblages during cooling from 625 to $525^\circ C$ (Fig. 4b). To illustrate the abundant growth of muscovite (Fig. 4b) the MKA and KFM projections (Fig. 7b, c) is used. For the same two $M_3$ parageneses, the modal proportion of muscovite increases, as the biotite stability field shrinks towards lower temperatures.

In summary, all three bulk compositions show different stages of the $PT$ evolution between $M_2$ and $M_3$ (Figs 6, 7 and 8). However, it is emphasized that, from Figs 6 and 7 alone, no unique $PT$ path can be inferred. Every $PT$ evolution from $M_2$ to $M_3$ that crosses the grey-shaded area in Fig. 6 may be possible. Some further constraints
Table 3: Average temperature calculations for M3 assemblages from the Nine-Mile Mine region

Average T calculated for chloritoid–staurolite assemblage (95BH19A)

<table>
<thead>
<tr>
<th>End-members:</th>
<th>mu</th>
<th>cel</th>
<th>mctd</th>
<th>fctd</th>
<th>mst</th>
<th>fst</th>
<th>phl</th>
<th>ann</th>
<th>q</th>
<th>H2O</th>
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<tr>
<td>Activities (a):</td>
<td>0.648</td>
<td>0.0201</td>
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<td>0.83</td>
<td>3.93e−4</td>
<td>0.545</td>
<td>0.0152</td>
<td>0.0886</td>
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<td>σ(ln a):</td>
<td>0.1</td>
<td>0.44</td>
<td>0.23</td>
<td>0.05</td>
<td>0.89</td>
<td>0.2</td>
<td>0.46</td>
<td>0.28</td>
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Independent reactions:

Calculated:

\[ T_{\text{cal}} \]

<table>
<thead>
<tr>
<th>( \sigma(T) )</th>
<th>d( T/dP )</th>
<th>ln(K)</th>
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<tbody>
<tr>
<td>0.01116</td>
<td>3.73</td>
<td></td>
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</tbody>
</table>

(1) \( 3\text{mctd} + \text{ann} = 3\text{fctd} + \text{phl} \)

Average temperatures calculated for a pressure series:

| P: | 3.0 | 4.0 | 5.0 | 6.0 | 7.0 | 8.0 |
| av. T: | 461 | 467 | 471 | 474 | 476 | 478 |
| σ(T): | 54 | 55 | 56 | 57 | 58 | 59 |

Average T calculated for garnet–staurolite assemblage (95BH19C)

<table>
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<th>fst</th>
<th>mst</th>
<th>mu</th>
<th>cel</th>
<th>phl</th>
<th>ann</th>
<th>q</th>
<th>H2O</th>
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</thead>
<tbody>
<tr>
<td>Activities (a):</td>
<td>1.32e−4</td>
<td>0.528</td>
<td>0.588</td>
<td>2.39e−4</td>
<td>0.65</td>
<td>0.0103</td>
<td>0.00967</td>
<td>0.0974</td>
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<td>1.0</td>
</tr>
<tr>
<td>σ(ln a):</td>
<td>0.81</td>
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<td>0.20</td>
<td>0.92</td>
<td>0.10</td>
<td>0.48</td>
<td>0.49</td>
<td>0.26</td>
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</table>

Independent reactions:

Calculated:

\[ T_{\text{cal}} \]

<table>
<thead>
<tr>
<th>( \sigma(T) )</th>
<th>d( T/dP )</th>
<th>ln(K)</th>
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<tbody>
<tr>
<td>0.01393</td>
<td>5.984</td>
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(1) \( \text{py} + \text{ann} = \text{alm} + \text{phl} \)

Average temperatures calculated for a pressure series:

| P: | 3.0 | 4.0 | 5.0 | 6.0 | 7.0 | 8.0 |
| av. T: | 496 | 460 | 467 | 471 | 474 | 476 |
| σ(T): | 54 | 55 | 56 | 57 | 58 | 59 |

All mineral analyses used for the calculations are listed in Table 2. Activities, \( a \), used here were calculated with the software ‘AX’ of Holland & Powell (unpublished software). av. T, average temperature of different independent reactions; \( K \), distribution coefficient; \( \sigma \), standard deviation; \( f \), fit parameter (Powell & Holland, 1988).

may be placed by the interpretation of pseudosections. Mineral assemblages of bulk composition 2 (e.g. 19A, 18B) show that the divariant equilibria st±mu±chl±as and st±bi±mu±as were crossed during cooling (Fig. 7c). These equilibria occur at very low pressures (Fig. 8a), probably below peak pressures of \( M_2 \), depending on the exact bulk composition. Correspondingly, some assemblages show that the divariant equilibria st±g±bi±mu and the st±g±chl±mu were crossed during cooling (Fig. 7a and b). These equilibria too occur at low pressures (Fig. 8b). Therefore, a cooling path from \( M_1 \) involving minor decompression may be indicated. For the divariant equilibria discussed above to occur at higher \( P \) than \( M_2 \), bulk compositions of rocks need to be much more magnesian than those observed. A cooling path involving minor decompression is also consistent with the rare occurrence of newly grown andalusite during cooling (Stevens, 1978) [Fig. 9a, path (ii)]. The \( M_1 \) peak may have been subsequently reached by compression and renewed heating from the stable geotherm [Fig. 9a, path (iii)].

The interpretation of the pseudosections may indicate that \( M_2 \) and \( M_3 \) are unrelated. However, the evidence remains ambiguous. Below, we further constrain the \( PT \) path by using independent field evidence.

**DISCUSSION**

The effect of MnO and accuracy of \( PT \) estimates

In KFMASH, temperature estimates for \( M_2 \) and \( M_3 \) and pressure estimates for \( M_2 \) are well constrained by the petrogenetic grid and pseudosections (Figs 6 and 8).
The absence of ctd–bi equilibria during $M_2$ is consistent with the grid topology as given in Fig. 6 and therefore with the topology suggested by Droop & Harte (1995).

The **PT path in relative time**

$M_2$ and $M_3$ metamorphic conditions are well constrained by average $PT$ calculations and interpretation of the petrogenetic grid, but the $PT$ path between the two events is not (Fig. 6). Of a series of possible $PT$ paths, we consider two important qualitatively different possibilities:

1. A direct cooling path, possibly involving minor decompression from $M_2$ to $M_1$ [Fig. 9a, open arrow: path [i]]. Such a $PT$ path was inferred by Clarke et al. (1987) for the Olary Block adjacent to the Broken Hill Block. (2) A cooling path involving minor decompression from $M_2$ [Fig. 9a, path [ii]] and independent, and much later, compression during $D_3$ towards $M_1$ [Fig. 9a, shaded arrows: path [iii]]. Apart from the interpretation of the pseudosection, which suggests possibility (2), the following information may help to discriminate between these possibilities:

   (1) $M_3$ occurred as static mineral growth and is separated from $M_2$ by at least one deformation episode, $D_3$ (Fig. 2). If the $PT$ path was continuous, then the similar pressures of $M_3$ and $M_4$ imply that the substantial shortening that occurred during $D_3$ did not cause any significant depth changes of the rocks. Alternatively, and more likely, $D_3$ may have caused crustal thickening substantially after cooling [Fig. 9a, path [iii]]. If the cooling path from $M_2$ was accompanied by exhumation, much later reburial and reheating towards, or beyond, $M_3$ pressures and temperature may have caused $M_1$ to occur at the same depth as $M_3$.

   (2) The regional distribution of $M_3$ isograds (Stevens et al., 1988, fig. 1a; this study, Fig. 1b) does not coincide with the regional distribution of $M_2$ isograds (Phillips, 1980; Figs 1 and 9c). This too, is most easily explained if $M_3$ was a thermal event independent of $M_2$.

   (3) Previous workers suggested that the $M_3$ assemblage is related to shear zones (e.g. Brown, 1983). They have stated that $M_3$ parageneses appear to occur proximal to, but not within, the many shear zones. These shear zones are generally 1–50 m in width, steeply dipping and extremely muscovite-rich. The geometry and shear sense of these zones is inconsistent with the regional geometry of $D_3$ deformation. Thus, if $M_3$ was related to shear zone development during $D_3$, it may be in an independent and much later event than the $M_3$-$S_3$ event.

These arguments indicate that $M_3$ and $M_4$ are likely to be unrelated. Indeed, previous attempts to correlate the retrograde metamorphism with the abundant geochronological data available (e.g. Stevens et al., 1990) have related peak and retrograde metamorphism to unrelated events.

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**Fig. 9.** (a) $M_1$ and $M_2$ metamorphic conditions of the Nine-Mile Mine region in comparison with the metamorphic peak conditions inferred for Round Hill (Powell & Downes, 1990) and possible $PT$ paths. (i) Continuous $PT$ path between $M_1$ and $M_2$. (ii) Preferred cooling path from $M_2$. (iii) Preferred $PT$ path during $M_1$. (iv) Possible extension of path (ii) if the inferred $M_1$ conditions in the Nine-Mile Mine region are below metamorphic peak conditions and if the metamorphic peak conditions were like those at Round Hill. (v) Prograde path before $M_3$ as suggested by abundant pseudomorphs of sillimanite after andalusite.

(b) A schematic north–south profile through the Broken Hill Block showing that $M_3$ and $M_4$ have different rates of grade change across the Block, possibly indicating that they are independent events.
Relationship of $M_2$ to peak conditions in adjacent areas

The $M_2$ conditions defined by the biotite–muscovite–sillimanite garnet assemblage are below the muscovite breakdown reaction (Fig. 6) and therefore substantially below the peak metamorphic grade suggested by Phillips (1980) (Fig. 1) (also Stevens, 1978; Brown, 1981). They are also much lower grade than the peak conditions of garnet–cordierite assemblages in the nearby Round Hill (Powell & Downes, 1990) (Fig. 9a and c). Metamorphic conditions in the Nine-Mile Mine region higher grade than those documented here are supported by abundant partial melts present in the area. Abundant partial melting is difficult to explain during $PT$ conditions as invoked here for $M_2$. Thus, it is likely that the peak temperatures in the Nine-Mile Mine region were originally higher than observed now, possibly during an $M_1$ event. However, retrogression was pervasive during $D_2$ so that little evidence of the preceding higher-grade conditions is preserved. It is interesting to note that the extent of retrogression just south of the study area is much less and higher-grade conditions are preserved. Possible $M_1$ metamorphic conditions for the Nine-Mile Mine rocks of 750°C at 5 kbar agree well with our interpretation of a $PT$ path involving minor decompression [Fig. 9a, path (iv)].

The $PT$ path in absolute time and its tectonic significance

$M_2$ is correlated with the ~1600 Ma event that is widely recognized as the metamorphic peak event in the Broken Hill district (e.g. Page & Laing, 1992; Ehlers & Nutman, 1997). $M_2$ is clearly younger than $M_1$ but its absolute age is unknown. If a continuous path from $M_1$ to $M_2$ is inferred, then it may be only some tens of million years younger than $M_1$. However, if formed during an independent event, as suggested here, then it may be significantly younger, for example, as young as the Grenvillian or Pan African events (Harrison & McDougall, 1981). Current geochronology will place further constraints on this event (Foster & K. Ehlers, unpublished data). Regardless of the interpretation of the $PT$ path, the cooling path from $M_2$ has all the characteristics of an externally heated terrain [see also Phillips & Wall (1981)]. This is suggested by the extremely perturbed $PT$ ratios during $M_2$ and the near-isobaric nature of the path for both possible interpretations.

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CONCLUSIONS

Meta-pelitic rocks from the central Broken Hill block show evidence for two metamorphic events, here termed $M_2$ and $M_3$. The first metamorphic event, $M_3$, occurred synchronous with an early deformation phase, $D_2$, and is characterized by the assemblage biotite–muscovite ± sillimanite ± garnet. This event occurred around 650°C and 4−5 kbar. The second metamorphic event, $M_2$, postdated a later deformation phase, $D_3$, as static overprint on all previous fabrics around 480°C and 5 kbar.

Meta-pelitic rocks record the retrograde path in spectacular detail by preserving a series of metastable equilibria. However, the topology of the petrogenetic grid allows a series of different trajectories across these equilibria. Two possibilities are of particular relevance: (1) the $PT$ path between $M_2$ and $M_3$ may be a straight $PT$ vector involving minor compression during cooling; (2) rocks may have cooled from $M_2$ with minor decompression and $M_3$ was attained much later during an independent thermal event involving some burial.

Interpretation of petrogenetic pseudosections suggests the latter of these possibilities. This interpretation is supported by: (1) $D_3$ causing substantial shortening throughout the region although $M_1$ pressures are only marginally higher than those of $M_2$; this is most easily explained if cooling from $M_2$ was accompanied by minor decompression and $M_3$ caused substantial shortening and that the peak temperatures in the Nine-Mile Mine were originally higher than observed now, possibly during an independent thermal event involving some burial.

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