

On the roles of deformation and fluid during rejuvenation of a polymetamorphic terrane: inferences on the geodynamic evolution of the Ruker Province, East Antarctica

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ABSTRACT Evaluating pressure–temperature (P – T) conditions through mineral equilibria modelling within an amphibolite facies polymetamorphic terrane requires knowledge of the fluid content of the rocks. The Archean–Palaeoproterozoic basement rocks of the Ruker Province, East Antarctica, preserve evidence of three metamorphic events (M1–M3). Of particular interest is the M3 event, which is constrained to the early Palaeozoic (*c.* 550–480 Ma). Evaluation of the tectonic setting during this time is important because the Ruker Province is located within a critical region with respect to models of Gondwana assembly. Structural evidence of the early Palaeozoic event is preserved as large (up to ~500 m wide) high strain zones that cut the orthogneiss–metasedimentary basement (Tingey Complex) of the Ruker Province. Rocks within these zones have been thoroughly recrystallized and preserve a dominant shear fabric and M3 mineral assemblages that formed at P – T conditions of 4.0–5.2 kbar and 565–640 °C. Distal to these zones, rocks preserve more complex petrographic relationships with S1 and S2 foliations, being incompletely overgrown by M3 retrograde assemblages. We show that the mineral assemblages preserved during the M3 event are highly dependent on the availability of fluid H_2O , which is strongly influenced by the structural setting (*i.e.* proximity to the high-strain zones). P – T structural and fluid flow constraints support a model of basin inversion during early Palaeozoic crustal rejuvenation in the Ruker Province.

Key words: East Antarctica; fluid flow; mineral equilibria modelling; southern Prince Charles Mountains; THERMOCALC.

INTRODUCTION

In rocks that have been previously metamorphosed, fluid absence during subsequent events can result in the limited development of new mineral assemblages during successive periods of metamorphism (Thompson, 1983; Guiraud *et al.*, 2001; Elmer *et al.*, 2006; Tenczer *et al.*, 2006). This can result in a metamorphic event being poorly preserved in the petrographic record. To further explore this, the roles of deformation and fluid access within a terrane that preserves evidence for several structural and metamorphic events is examined. The approach taken uses mineral equilibria modelling to examine the evolution of available fluid H_2O during successive phases of amphibolite grade metamorphism. This illustrates phase equilibria during amphibolite facies metamorphism and its effects on the availability of H_2O during successive metamorphic events. In addition, the relationship between structural

controls leading to fluid infiltration from an external source, and the rejuvenation of polymetamorphic terranes, are also explored. The Ruker Province of East Antarctica is used to further elucidate these relationships (Fig. 1a,b).

Studying the metamorphic evolution of the Ruker Province has further significance in that it was situated within a key location with respect to Gondwana assembly. Preliminary work from the Ruker Province has identified three metamorphic events M1–M3 (after the divisions in Mikhalsky *et al.*, 2001), in which P – T conditions are poorly constrained. Of particular interest is the M3 event, which represents a dominant phase of orogenic activity that coincides with the assembly of Gondwana (*c.* 550–480 Ma; Tingey, 1991; Boger *et al.*, 2001; Phillips *et al.*, 2007). There is disagreement over the tectonic setting during this event, with models of continent–continent collision (Boger *et al.*, 2001) or basin inversion (Phillips *et al.*, 2007) both invoked. An accurate evaluation of orogenic conditions (P – T , deformation and fluid conditions) during this time period is critical if an agreeable tectonic model for the region is to be realised.

Within this paper we present new data that constrains P – T , deformation and fluid conditions (M_{H_2O})

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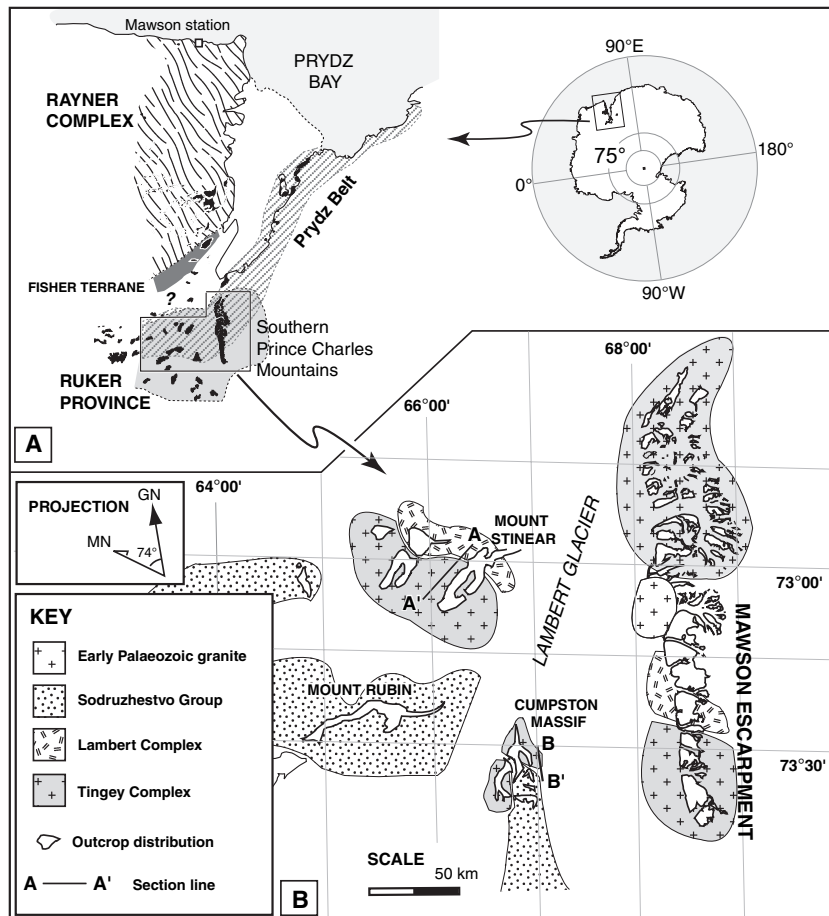


Fig. 1. Maps of the study area and the main lithotectonic subdivisions. (a) Location of the Prince Charles Mountains-Prydz Bay region and key tectonic components. The focus of this paper is the Ruker Province, which is within the southern Prince Charles Mountains. (b) Stylized outcrop map of the southern Prince Charles Mountains and the main lithotectonic divisions. Section lines are for cross-sections A-A' (Mount Stinear) and B-B' (Cumpston Massif) presented in Fig. 2.

for the evolution of the polymetamorphic Ruker Province (Phillips *et al.*, 2006). This allows inferences to be made on the roles of fluid and deformation (White & Clarke, 1997) in determining the evolution, preservation and distribution of metamorphic mineral assemblages during successive orogenic events. The resultant structural, P - T and fluid flow constraints will be tested against the current tectonic models.

BACKGROUND AND REGIONAL GEOLOGY

The Ruker Province is located in the southern Prince Charles Mountains of East Antarctica, and is exposed as a series of isolated ranges (Fig. 1a,b). The significance of studying the Ruker Province is that it is located to the south of the 550–500 Ma aged Prydz Belt (Fig. 1a), which some authors argue, represents the site of collision between India (and the attached Rayner Complex) and East Antarctica during Gondwana assembly (Boger *et al.*, 2001). In sharp contrast, recent studies have alluded to the possibility that the Prydz Belt may have formed during the inversion of a major intra-cratonic Neoproterozoic basin, which was driven by far-field tectonic stress (Phillips *et al.*, 2007). These two conflicting tectonic scenarios would invoke con-

trasting geodynamic conditions within the Ruker Province, which should be apparent in the structural-metamorphic record preserved.

Current stratigraphic/structural divisions of the Ruker Province are summarized in Table 1, and discussed herein. The Ruker Province comprises three main lithotectonic units that include: (1) the Archean-Palaeoproterozoic (*c.* 3200–2300 Ma) Tingey Complex, (2) the Palaeoproterozoic-Mesoproterozoic (*c.* 2100–1000(?) Ma) Lambert Complex and (3) the

Table 1. Summary of the main geological events and structural divisions of the Ruker Province.

Ma	Event	
c. 550–480	D3/M3	Emplacement of pegmatite stocks and granite
		Folding of the Sodruzhestvo Group
		HSZ in the Tingey Complex
< 900		Sodruzhestvo Group deposited
< 1800		Deposition of the Lambert Group
c. 2100	?D2/M2?	Orthogneiss emplaced
< 2500		Deposition of the Ruker Group
< 2800		Deposition of the Stinear Group
		Isoclinal folds
< 3170	D1/M1	Deposition of the Menzies Group
3190–3170		Mawson orthogneiss emplaced

HSZ, high-strain zones.

Neoproterozoic (*c.* 950–600 Ma) Sodruzhestvo Group. The main focus of the present study is the basement (Tingey Complex) and cover (Sodruzhestvo Group) rocks of the Ruker Province.

Tingey Complex

The oldest exposed unit is the Mawson orthogneiss, which was emplaced between *c.* 3190 and 3170 Ma (U-Pb – SHRIMP zircon geochronology; Boger *et al.*, 2006). Deposition of the Menzies Group occurred after emplacement of this unit, which is based on U-Th-Pb dating of detrital zircon (LA-ICPMS; Phillips *et al.*, 2006). A tectonothermal event that is constrained by U-Pb (SHRIMP) geochronology on magmatic zircon from partial melts within the Mawson orthogneiss is suggested to have occurred at *c.* 2800–2780 Ma (Boger *et al.*, 2006). This event was probably responsible for the development of F1 isoclinal folds within the Archean orthogneiss and a bedding parallel foliation within the metasedimentary units of the Menzies Group (Fig. 2a; Phillips *et al.*, 2005). Deposition of the quartzite-dominated Stinear Group post-dates the D1 event, and contains a pronounced *c.* 2800–2780 Ma aged detrital zircon signature (U-Th-Pb – LA-ICPMS; Phillips *et al.*, 2006). D2 folds with shallowly plunging hinges characterize the main regional deformation feature of the Tingey Complex, with examples exposed at Mount Stinear (Figs 1b, 2a & 3a). Deposition of the Ruker Group occurred after this event (*c.* 2500 Ma), constrained by dating of detrital zircon (U-Th-Pb – LA-ICPMS; Phillips *et al.*, 2006).

Overprinting all earlier structural features of the Tingey Complex are prominent D3 high-strain zones (HSZ) that are exposed at Mount Stinear, and range from 5 to 500 m in width (Figs 1b, 2a & 3b). These zones generally trend to the north-west, are characterized by a strong stretching lineation that plunges moderately to the north-west and have formed proximal to lithological contacts between the main stratigraphic units (Phillips *et al.*, 2005; Fig. 3b). This most recent phase of tectonic activity has been constrained to 520–480 Ma by the systematics of argon closure within hornblende, biotite and white mica from the metasedimentary rocks (Phillips *et al.*, 2007).

Lambert Complex

The Lambert Complex (Fig. 1b) comprises orthogneiss and metasedimentary units that are discriminated from the Ruker Province by different populations of U-Pb magmatic, metamorphic (SHRIMP) and detrital zircon ages (LA-ICPMS) (*c.* 2600, 2100–1800, 1600 and 500 Ma; Boger *et al.*, 2001; Phillips *et al.*, 2006). Emplacement of the orthogneiss has been interpreted to have occurred at *c.* 2100 Ma, which was followed by deposition of the overlying metasedimentary rocks (Lambert Group) after *c.* 1800 Ma (Phillips *et al.*, 2006). The peak preserved metamorphic mineral

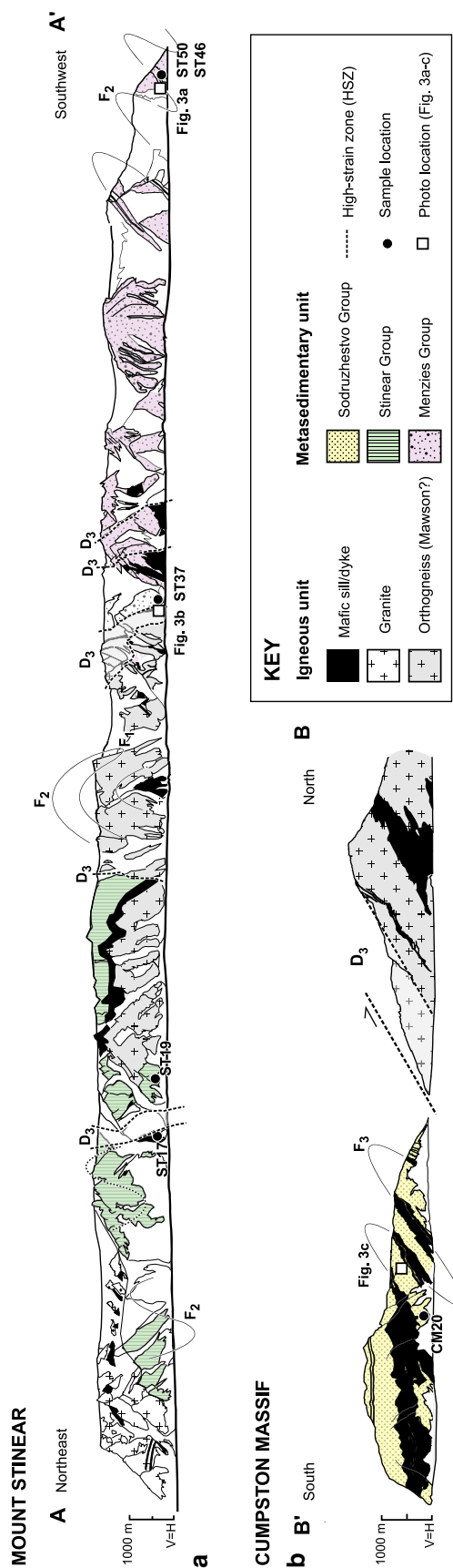


Fig. 2. Cross-sections showing the main structural features of (a) the Tingey Complex – exposed at Mount Stinear and (b) the Sodrzhestvo Group – exposed at Cumpston Massif. Samples locations are shown. Note the locations of photographs shown in Fig. 3.

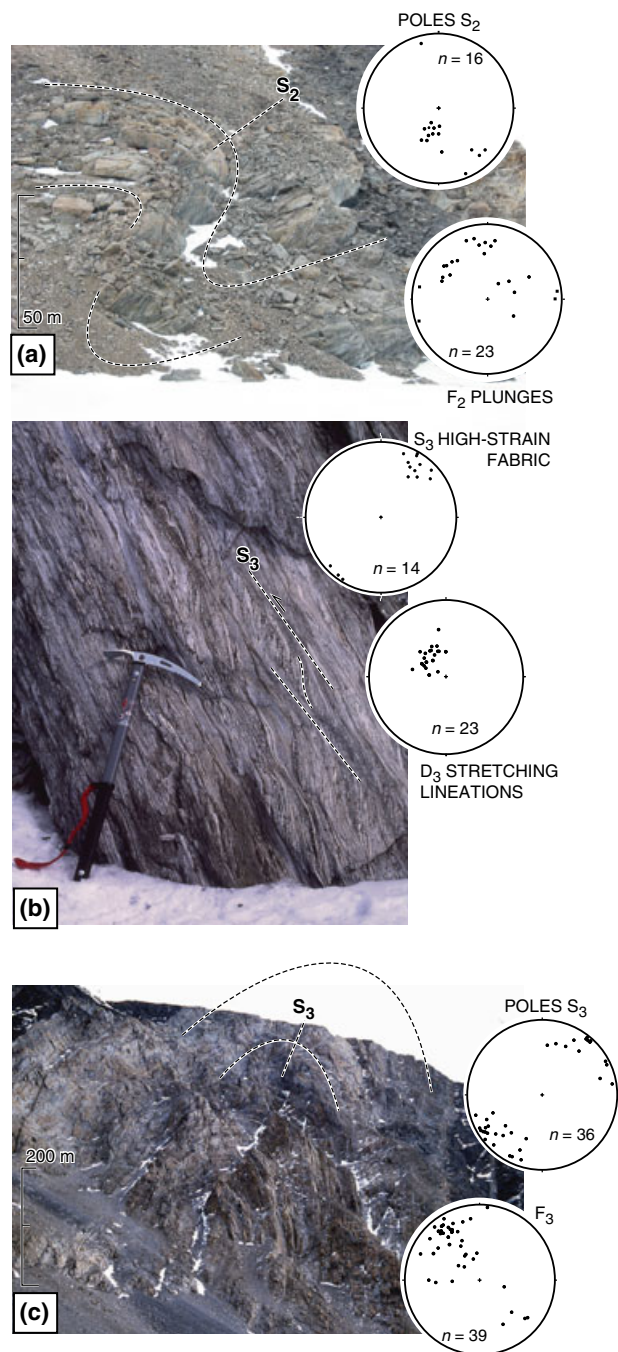


Fig. 3. Field photographs and structural data of the key meso-scale observations. (a) D2 regional fold structures at Mount Stinear – Tingey Complex. (b) D3 high-strain zone that juxtaposes basement orthogneiss with Archean metasedimentary rocks at Mount Stinear – Tingey Complex. (c) D3 fold structure at Cumpston Massif – Sodruzhestvo Group. Photograph locations are shown on cross-sections in Fig. 2.

assemblages within the metasedimentary units formed at conditions between 650–700 °C and 6.0–7.0 kbar (Boger & Wilson, 2005). This event has been associated with folding of the orthogneiss/metasedimentary package into upright, long wavelength structures that

plunge shallowly to the north-west and west (Boger & Wilson, 2005). Constrained by the systematics of argon closure within biotite, final cooling of the Lambert Complex below ~300 °C (Harrison *et al.*, 1985) occurred at 486 ± 8 Ma (Phillips *et al.*, 2007).

Sodruzhestvo Group

The youngest unit of the Ruker Province is a thick (up to 6 km) package of Neoproterozoic metasedimentary rocks, termed the Sodruzhestvo Group (Figs 1b & 2b; Kamenev *et al.*, 1993; Phillips *et al.*, 2006).

Based on detrital zircon data (U-Th-Pb – LA-IC-PMS) deposition of this unit occurred after *c.* 950 Ma (Phillips *et al.*, 2006). A phase of extension during the Neoproterozoic, which was attributed to the development of regional basins, has been argued on the basis of the deposition of this unit and the localization of thick mafic sills (up to 500 m) within the base of the exposed metasedimentary succession (Phillips *et al.*, 2005, 2006). A single phase of shortening has folded the metasedimentary succession into upright to locally reclined, north-west-trending structures that have a pervasive axial planar S1 foliation (Figs 2b & 3c; Phillips *et al.*, 2005). Based on the middle Neoproterozoic depositional age of the Sodruzhestvo Group and the systematics of argon closure within S1 white mica at 503–480 Ma, the D1 event is constrained to the early Palaeozoic (Phillips *et al.*, 2007). Deformation and metamorphism of the Sodruzhestvo Group actually represents the D3 event within the regional structural division, and is referred to as S3/M3 herein (Table 1).

PETROGRAPHY

Textural relationships between metamorphic mineral growth and structural features constrain the relative timing of metamorphism with respect to deformation events. Textural relationships within metasedimentary rocks of the basement Tingey Complex and cover Sodruzhestvo Group have been used to constrain the timing of key tectonic events. Evidence for an early metamorphic event, M1, is locally preserved as a bedding parallel foliation evident only within the Tingey Complex. Two later metamorphic events (M2 and M3) are identifiable in the mineral record of the Tingey Complex, which contrasts with evidence for only one event (M3) preserved in the Sodruzhestvo Group.

Tingey Complex

Samples from the Tingey Complex were selected for comparison of metamorphic mineral assemblages proximal and distal to the D3 HSZ. Samples ST46 and ST50 are located distal to any HSZ (Fig. 2a) and

preserve the effects of the early orogenic history of the region. Conversely, samples ST17, ST19 and ST37 were selected because of proximity to the D3 HSZ (Fig. 2a), and therefore preserve dominantly the effects of M3 metamorphism/deformation.

Distal to HSZ – M2

Samples ST50 and ST46 ($73^{\circ}07.752'S/66^{\circ}10.230'E$), from the southernmost tip of Mount Stinear (Figs 1b & 2a), are part of an F2 asymmetrically folded package of biotite-rich schist, amphibole–garnet-bearing metabasite and quartzite (Figs 2a & 3a). The S2 foliation in these samples is parallel to the axial plane of the regional fold structures shown in Fig. 3a. The metapelite sample, ST50, is composed of a recrystallized polygonal quartz groundmass along with anhedral grains of biotite, muscovite, kyanite and clinozoisite. Preferably-oriented muscovite and biotite laths define the earliest identifiable foliation, S1. These minerals contain numerous quartz inclusions and are crenulated, leading to an S2 foliation that is defined by a muscovite–biotite–kyanite-rich assemblage. The porphyroblasts of kyanite have been overgrown by aggregates of fine-grained white mica (Fig. 4a). Porphyroblasts of clinozoisite overgrow all early assemblages (Fig. 4b), and have also been partially replaced by fine-grained white mica.

The metabasite sample, ST46, has the assemblage quartz, chlorite, biotite, garnet and amphibole. Early S1 biotite contains numerous quartz inclusions, and is wrapped by a S2 chlorite-rich fabric. The S1 biotite has been partially pseudomorphed by chlorite along grain boundaries. The dominant S2 fabric is defined by aggregates of amphibole and garnet along with chlorite; the latter wraps around, and is also overgrown by the amphibole (Fig. 4c) and garnet.

Proximal to HSZ – M3

Sample ST17 ($73^{\circ}03.856'S/66^{\circ}18.714'E$) is from a sheared metaconglomerate that is located proximal to a north-west-trending D3 HSZ at Mount Stinear (Fig. 2a). Deformation of the rock is characterized on the mesoscale by the highly elongate nature of the quartz-rich clasts from the proto-conglomerate. At the microscale, the polygonal quartz-rich matrix within both the clasts and the enclosing sedimentary matrix have been highly recrystallized. The sedimentary matrix exhibits a moderately strong biotite ($\pm$ chlorite)-rich foliation (S3) that occurs along the quartz grain boundaries. The quartz–biotite–chlorite assemblage is overgrown by a garnet–staurolite–magnetite assemblage that is inferred to be the peak metamorphic assemblage (Fig. 5a). Fine-grained euhedral garnet in the quartz-rich clasts contrasts with larger, anhedral grains that occur within the sedimentary matrix. Garnet also commonly occurs as inclusions within the larger staurolite aggregates, which additionally contain

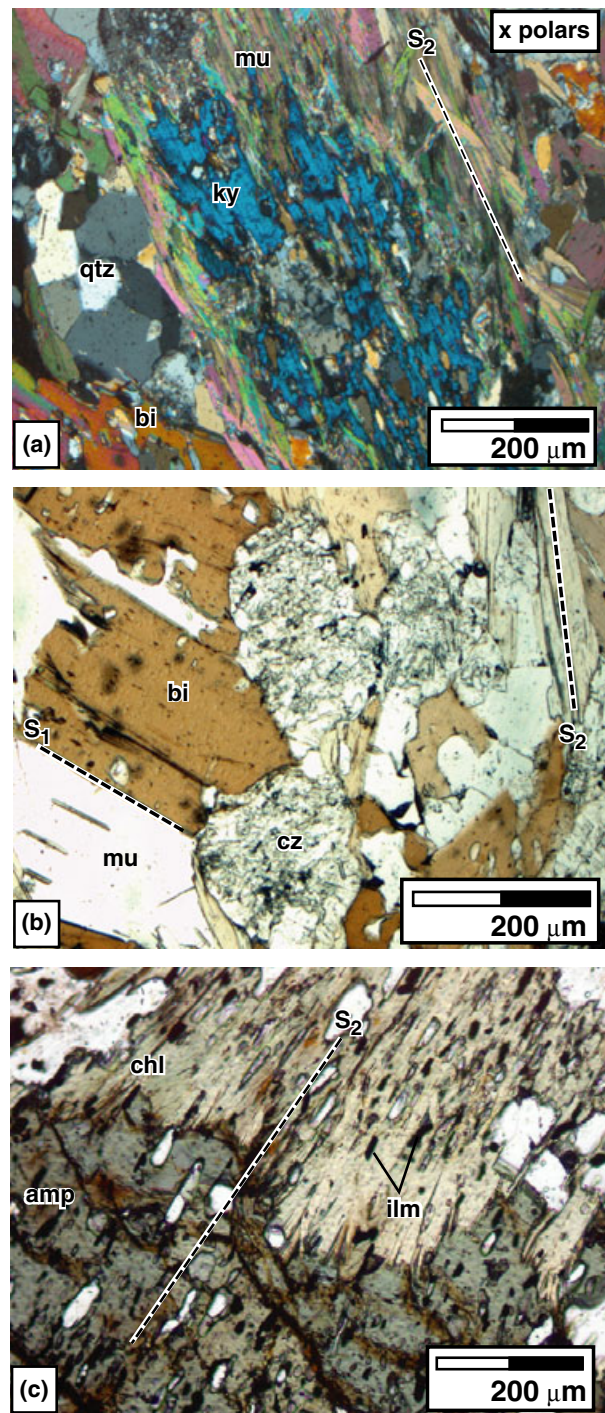


Fig. 4. Photomicrographs of key petrographic observations with respect to the M2 assemblages from Mount Stinear – Tingey Complex. (a) Kyanite–white mica porphyroblast aligned with the S2 foliation – sample ST50 (polars crossed). (b) M3 clinozoisite overprinting S1 biotite and S2 muscovite – sample ST50. (c) Syn- to post-tectonic amphibole and chlorite defining the S2 foliation – sample ST46.

numerous quartz inclusions. Thin fractures that are generally oblique to the orientation of the main foliation cut both the garnet and staurolite aggregates.

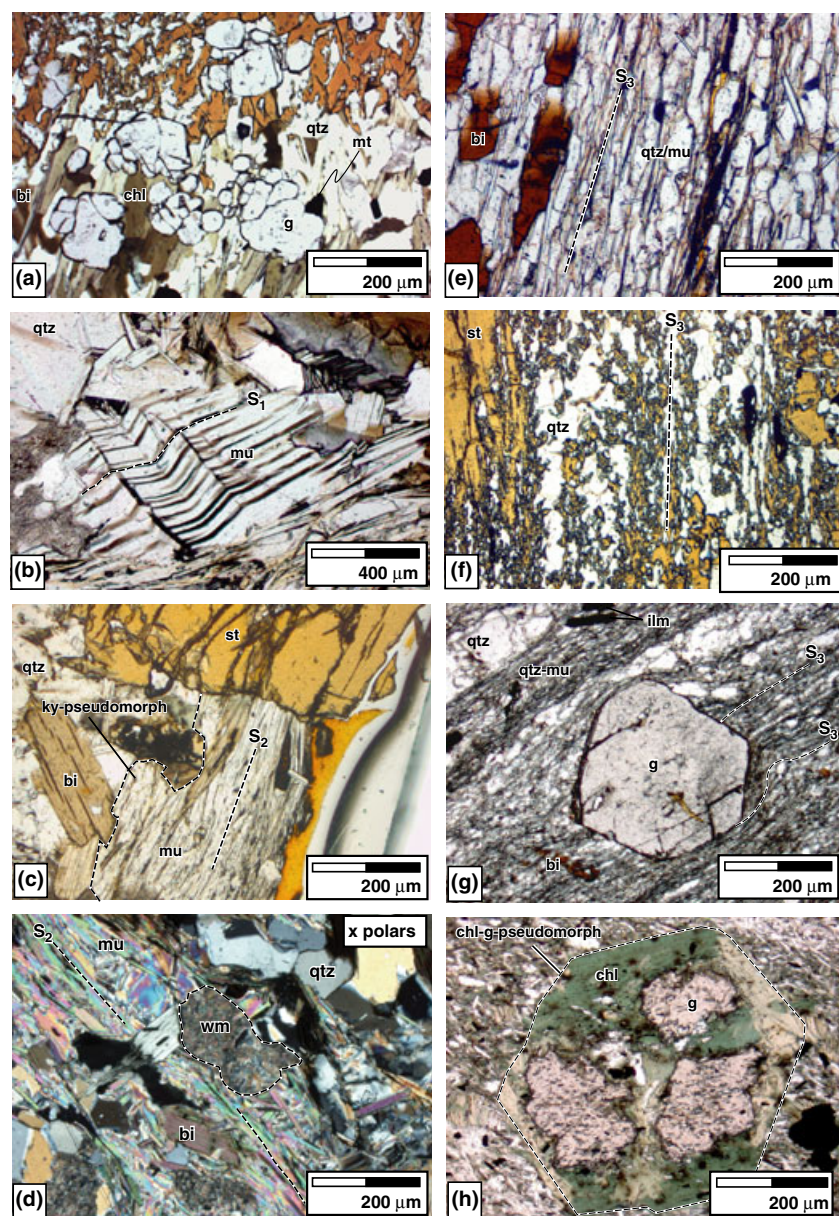


Fig. 5. Photomicrographs of key petrographic observations with respect to the M3 assemblages at Mount Stinear and Cumpston Massif. (a) M3 staurolite-garnet overprinting chlorite-biotite foliation. (b) Kinked S1 muscovite – sample ST19. (c) M3 staurolite and retrograde fine-grained white mica that has pseudomorphed kyanite – sample ST19. (d) Late-stage fine-grained white mica clots overprinting the S2 foliation. (e) S3 tectonic fabric within HSZ – sample ST37. (f) M3 staurolite intergrown with the S3 fabric within HSZ – sample ST37. (g) M3 syn- to post-tectonic garnet – sample CM20. (h) Chlorite pseudomorphs around M3 garnet – sample CM45.

Within this sample there is no evidence for the earlier S1 or S2 foliations.

Sample ST19 (73°04.029'S/66°18.300'E) is from a package of metapelite and metapsammite located ~200 m from a sub-vertical D3 HSZ (Fig. 2a). The matrix of the sample is polygonal fine-grained quartz along with muscovite, biotite, kyanite and staurolite. S1 muscovite and biotite are kinked (Fig. 5b), and are overgrown by elongate aggregates of muscovite and kyanite laths defining S2. Euhedral M3 staurolite overgrows the S2 foliation (Fig. 5c), and also contains S2 parallel quartz inclusions. Rounded clots of fine-grained white mica overprint all assemblages (Fig. 5d).

Sample ST37 (73°05.801'S/66°14.266'E) is from within a D3 HSZ that is located along the contact between orthogneiss and the Menzies Group (Fig. 2a). The matrix of the rock is fine-grained, highly recrystallized, and is made up of polygonal grains of quartz and plagioclase. Intergrown with the matrix is fine-grained white mica that defines the S3 shear fabric (Fig. 5e). S3 biotite, ilmenite and staurolite preserve distinctive sievelike textures, and comprise the remainder of the sample (Fig. 5f). The anhedral staurolite and biotite porphyroblasts overgrow the S3 foliation, which is evident by the numerous S3 parallel quartz inclusions within these minerals. Within this sample there is no evidence for the earlier S1 or S2 foliations.

Sodruzhestvo Group

Cover succession – M3

Sample CM20 (73°36.064'S/66°58.501'E) is from a folded package of metapelite located within the lower stratigraphic levels of the Sodruzhestvo Group at Cumpston Massif (Fig. 2b). The sample consists of a fine-grained quartz-muscovite ± plagioclase matrix and larger grains of euhedral garnet (Fig. 5g), ilmenite, biotite and minor chlorite. Elongate domains of fine-grained muscovite define the dominant S3 fabric in the rock. The M3 garnet primarily overgrows the matrix, yet there is also evidence for the S3 fabric wrapping the larger euhedral grains (Fig. 5g). Chlorite locally overgrows biotite and ilmenite. Based on the textural relationships between the S3 fabric and the peak metamorphic garnet, a syn- to post-peak relationship between deformation and peak metamorphism is inferred (Fig. 5g).

Rocks located in a higher stratigraphic position within the Sodruzhestvo Group (CM45; 73°41.231'S/66°57.047'E) show a pronounced chlorite overprint that pseudomorphs the peak assemblages described in sample CM20. In these samples chlorite completely replaces biotite and partially replaces garnet (Fig. 5h). This relationship contrasts with the lower stratigraphic levels, where pristine euhedral garnet is preserved (cf. Fig. 5g,h).

MINERAL CHEMISTRY

Mineral compositions were acquired using a Cameca SX50 electron microprobe at The University of Melbourne Joint Electron Microprobe (JEM) Facility. Analyses were undertaken using wavelength dispersive spectrometers with an accelerating voltage of ~15 kV, a current of ~35 nA and a beam diameter of ~3 µm. Mineral analyses are presented in Table 2.

S2 garnet in the metabasite sample ST46 is almandine-rich with $\text{alm}_{73}\text{sps}_{13}\text{py}_{5}\text{gr}_9$ where $\text{alm} = 100 \times \text{Fe}/(\text{Fe} + \text{Mg} + \text{Mn} + \text{Ca})$, $\text{sps} = 100 \times \text{Mn}/(\text{Fe} + \text{Mg} + \text{Mn} + \text{Ca})$, $\text{py} = 100 \times \text{Mg}/(\text{Fe} + \text{Mg} + \text{Mn} + \text{Ca})$ and $\text{gr} = 100 \times \text{Ca}/(\text{Fe} + \text{Mg} + \text{Mn} + \text{Ca})$. No compositional zoning is evident from rim-core transects across individual grains (Table 2). S3 garnet in sample ST17 is also almandine-rich with $\text{alm}_{71}\text{sps}_{20}\text{py}_3\text{gr}_6$. Similar to garnet in sample ST46, rim-core transects showed no compositional zonation (Table 2). S3 garnet from sample CM20 is also almandine-rich, yet shows distinct zoning with more pyrope and grossular-rich cores. The chemistry of these grains ranges from $\text{alm}_{76}\text{sps}_{13}\text{py}_2\text{gr}_8$ to $\text{alm}_{60}\text{sps}_{5}\text{py}_{17}\text{gr}_{19}$ for rim to core respectively.

S2 amphibole in the sample ST46 is ferro-tschermakite (Leake *et al.*, 1997). S2 biotite from the same sample has $X_{\text{Fe}} [\text{Fe}/(\text{Fe} + \text{Mg})] = 0.48\text{--}0.50$. This contrasts with lower X_{Fe} in S2 biotite from samples ST50 ($X_{\text{Fe}} = 0.24\text{--}0.27$) and ST37 ($X_{\text{Fe}} = 0.42\text{--}0.44$).

All varieties of S2 biotite have relatively low Ti content that range between 0.13 and 0.20 Ti per formula unit (pfu) for 22 oxygens. S3 biotite from sample ST17 has $X_{\text{Fe}} = 0.40\text{--}0.45$ and 0.15–0.25 Ti pfu, which is similar to sample ST37, where $X_{\text{Fe}} = 0.40\text{--}0.46$ and 0.17–0.22 Ti pfu. S3 biotite from sample CM20 has higher Fe with $X_{\text{Fe}} = 0.55\text{--}0.64$, and similar Ti contents with 0.16–0.22 Ti pfu.

Muscovite-defining S1 in sample ST19 is approximately 13% paragonite, which compares with 10% for that defining S2 in the same sample. S2 white mica in sample ST50 has a similar paragonite content of 12%. Plagioclase feldspar in sample ST37 has $X_{\text{an}} [\text{Ca}/(\text{Ca} + \text{Na})] = 0.17$. Ilmenite in samples ST46 and CM20 contains between 0.2 and 1.5% pyrophanite. FeO and Fe₂O₃ content in magnetite in sample ST17 is between 2.85 and 2.93 pfu (3 oxygens).

MINERAL EQUILIBRIA MODELLING

Pseudosections were constructed for several bulk compositions to constrain *P–T* conditions of M2 and M3. Bulk rock compositions are normalized whole rock XRF data, and are presented in figure captions. The pseudosections were constructed using THERMOCALC 3.23 (Powell & Holland, 1988) and the internally consistent data set of Holland & Powell (1998) (data set tcds55, file created 22/11/2003). Calculations for amphibole-bearing rocks were undertaken in NCKFMASH. The following *a–x* models were used: amphibole (Dale *et al.*, 2005); garnet (Dale *et al.*, 2000); muscovite/paragonite (Coggon & Holland, 2002); biotite (Powell & Holland, 1999); chlorite (Holland *et al.*, 1998) and plagioclase (Holland & Powell, 2003). Calculations for metapelitic compositions were undertaken in MnNCKFMASHTO, using the following *a–x* relationships: garnet (White *et al.*, 2005); biotite (White *et al.*, 2005); chlorite (Mahar *et al.*, 1997; Holland *et al.*, 1998); muscovite/paragonite (Coggon & Holland, 2002); staurolite (Mahar *et al.*, 1997); plagioclase (Holland & Powell, 2003); epidote (Holland & Powell, 1998); ilmenite (White *et al.*, 2005); and magnetite (White *et al.*, 2000).

P–T pseudosections

Distal to HSZ – M2

Sample ST46. The *P–T* pseudosection for this sample is characterized by a series of mineral assemblage fields with steep boundaries in *P–T* that represent the transition from chlorite–biotite to biotite–garnet–amphibole-bearing mineral assemblages (Fig. 6a). The identification of biotite-defining S1 in this sample indicates that the earlier metamorphic event (M1) occurred at conditions above that of the biotite-in line in Fig. 6a. The development of the strong chlorite–amphibole–garnet-rich fabric is attributed to the M2 event and can be constrained to *P–T* conditions

Table 2. Electron microprobe data for individual mineral analyses.

Tingey Complex – distal to D3 high-strain zones															
	Sample ST6			Sample ST50				Sample ST46							
	ky	mu (S ₂)	bi (S ₂)	ky	mu (S ₂)	bi (S ₂)	cz	bi (S ₂)	ilm	amp	g transect				
SiO ₂	36.63	48.36	37.05	37.24	49.15	39.13	45.03	36.53	0.01	40.56	37.26	37.32	37.27	36.53	37.21
TiO ₂	0.02	0.30	1.82	0.00	0.68	1.24	0.02	1.70	51.91	0.48	0.08	0.02	0.11	0.06	0.08
Al ₂ O ₃	62.90	38.45	19.87	62.94	36.56	19.65	34.73	16.90	0.04	17.07	21.24	21.29	21.27	20.99	21.14
Cr ₂ O ₃	0.03	0.00	0.05	0.18	0.21	0.22	0.00	0.02	0.00	0.00	0.00	0.01	0.00	0.00	0.01
FeO	0.13	0.57	15.76	0.13	0.74	10.12	0.02	19.79	44.31	18.74	33.45	33.18	32.90	32.74	32.90
MnO	0.00	0.00	0.14	0.00	0.00	0.26	0.03	0.09	1.44	0.18	2.23	2.15	2.69	2.39	2.33
MgO	0.02	0.47	12.32	0.00	0.98	16.39	0.00	11.02	0.03	7.00	3.41	3.43	3.30	3.36	3.34
CaO	0.01	0.00	0.01	0.01	0.01	0.00	18.39	0.00	0.00	10.08	3.11	3.05	3.21	3.09	3.25
Na ₂ O	0.00	1.99	0.28	0.00	0.90	0.19	1.06	0.12	0.00	1.67	0.00	0.01	0.02	0.02	0.03
K ₂ O	0.00	7.92	9.18	0.00	9.58	9.64	0.02	9.50	0.00	0.30	0.01	0.00	0.01	0.01	0.00
Total	99.74	98.06	96.48	100.52	98.81	96.84	99.30	96.48	96.48	96.48	96.48	96.48	96.48	96.48	96.48
Oxygen	5	22	22	5	22	22	13	22	3	23	12	12	12	12	12
Si	0.99	6.14	5.44	1.00	6.24	5.57	3.40	5.55	0.00	6.18	2.97	2.98	2.97	2.96	2.98
Ti	0.00	0.03	0.20	0.00	0.07	0.13	0.00	0.19	1.01	0.05	0.00	0.00	0.01	0.00	0.00
Al	2.01	5.75	3.44	1.99	5.47	3.30	3.09	3.03	0.00	3.07	2.00	2.00	2.00	2.01	1.99
Cr	0.00	0.00	0.01	0.00	0.02	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe	0.00	0.06	1.94	0.00	0.08	1.20	0.00	2.52	0.95	2.39	2.23	2.22	2.19	2.22	2.20
Mn	0.00	0.00	0.02	0.00	0.00	0.03	0.00	0.01	0.03	0.02	0.15	0.15	0.18	0.16	0.16
Mg	0.00	0.09	2.70	0.00	0.19	3.48	0.00	2.50	0.00	1.59	0.41	0.41	0.39	0.41	0.40
Ca	0.00	0.00	0.00	0.00	0.00	0.00	1.49	0.00	0.00	1.65	0.27	0.26	0.27	0.27	0.28
Na	0.00	0.49	0.08	0.00	0.22	0.05	0.15	0.03	0.00	0.49	0.00	0.00	0.00	0.00	0.00
K	0.00	1.28	1.72	0.00	1.55	1.75	0.00	1.84	0.00	0.06	0.00	0.00	0.00	0.00	0.00
Total	3.00	13.84	15.54	3.00	13.84	15.54	8.13	15.68	1.99	15.50	8.03	8.02	8.02	8.03	8.02
Tingey Complex – proximal to D3 high-strain zones															
	Sample ST17							Sample ST19							
	st	bi (S ₃)	mag	g transect				ky	st	mu (S ₁)		mu (S ₂)			
SiO ₂	27.58	36.16	0.00	37.71	37.37	37.19	37.39	37.70	37.63	36.80	26.72	49.59	51.57		
TiO ₂	0.55	1.47	0.04	0.01	0.02	0.02	0.00	0.00	0.02	0.01	0.55	0.68	0.48		
Al ₂ O ₃	52.38	18.59	0.22	21.51	21.49	21.32	21.28	21.37	21.40	62.65	54.31	37.34	34.65		
Cr ₂ O ₃	0.11	0.15	1.71	0.11	0.06	0.02	0.10	0.02	0.01	0.16	0.17	0.15	0.20		
FeO	14.84	17.01	90.08	32.83	32.72	32.55	32.75	32.90	32.07	0.27	12.62	1.25	1.38		
MnO	0.06	0.01	0.00	1.56	1.69	1.73	1.71	1.69	1.62	0.00	0.78	0.02	0.04		
MgO	2.71	12.80	0.00	5.04	4.81	4.84	4.86	4.84	4.61	0.00	2.15	0.77	1.56		
CaO	0.01	0.00	0.00	2.01	2.19	2.46	2.52	2.43	3.20	0.02	0.01	0.00	0.00		
Na ₂ O	0.00	0.55	0.04	0.01	0.01	0.01	0.00	0.00	0.02	0.00	0.01	0.93	0.73		
K ₂ O	0.01	8.00	0.00	0.01	0.00	0.02	0.00	0.00	0.00	0.00	0.01	9.68	9.68		
Total	96.48	96.48	96.48	96.48	96.48	96.48	96.48	96.48	96.48	96.48	96.48	96.48	96.48		
Oxygen	24	22	3	12	12	12	12	12	12	5	24	22	22		
Si	4.01	5.43	0.00	2.98	2.97	2.96	2.97	2.98	2.98	1.00	3.89	6.21	6.46		
Ti	0.06	0.17	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.06	0.06	0.05		
Al	8.97	3.29	0.01	2.00	2.01	2.00	1.99	1.99	2.00	2.00	9.31	5.51	5.11		
Cr	0.01	0.02	0.05	0.01	0.00	0.00	0.01	0.00	0.00	0.00	0.02	0.02	0.02		
Fe	1.80	2.14	2.90	2.17	2.17	2.17	2.17	2.17	2.12	0.01	1.54	0.13	0.14		
Mn	0.01	0.00	0.00	0.10	0.11	0.12	0.11	0.11	0.11	0.00	0.10	0.00	0.00		
Mg	0.59	2.87	0.00	0.59	0.57	0.58	0.57	0.57	0.54	0.00	0.47	0.14	0.29		
Ca	0.00	0.00	0.00	0.17	0.19	0.21	0.21	0.21	0.27	0.00	0.00	0.00	0.00		
Na	0.00	0.16	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.23	0.18		
K	0.00	1.53	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.55	1.55		
Total	15.45	15.60	2.97	8.02	8.03	8.04	8.04	8.03	8.02	3.00	15.39	13.85	13.79		
Tingey Complex – proximal to D3 HSZ															
	Sample ST37				Sodruzhestvo Group										
	st	bi (S ₂)	bi (S ₃)	fsp	bi (S ₃)	ilm	g transect								
SiO ₂	27.57	36.90	37.02	66.11	34.40	0.01	36.78	36.38	36.59	36.45	36.51				
TiO ₂	0.49	1.61	1.69	0.00	1.64	53.93	0.02	0.08	0.15	0.03	0.00				
Al ₂ O ₃	54.08	19.86	19.71	22.78	18.61	0.00	20.94	20.81	20.78	20.80	20.97				
Cr ₂ O ₃	0.14	0.12	0.12	0.03	0.04	0.01	0.06	0.00	0.03	0.01	0.04				
FeO	13.11	16.31	16.21	0.03	22.31	44.79	35.23	31.61	25.43	31.48	36.41				
MnO	0.29	0.12	0.10	0.01	0.05	0.19	0.95	3.78	7.15	3.81	0.98				
MgO	2.42	12.07	12.27	0.01	8.22	0.05	1.71	0.91	0.57	0.92	1.80				
CaO	0.00	0.00	0.00	3.54	0.00	0.00	4.55	6.21	9.31	6.20	3.09				
Na ₂ O	0.01	0.39	0.39	9.53	0.14	0.01	0.05	0.04	0.01	0.06	0.04				

Table 2. *Cont'd*

	Tingey Complex – proximal to D3 HSZ				Sodruzhestvo Group						
	Sample ST37				Sample CM20						
	st	bi (S ₂)	bi (S ₃)	fsp	bi (S ₃)	ilm	g transect				
K ₂ O	0.01	8.49	8.75	0.03	8.23	0.01	0.00	0.00	0.00	0.00	0.00
Total	96.48	96.48	96.48	96.48	96.48	96.48	96.48	96.48	96.48	96.48	96.48
Oxygen	24	22	22	8	22	3	12	12	12	12	12
Si	3.97	5.45	5.45	2.85	5.39	0.00	2.97	2.96	2.96	2.97	2.97
Ti	0.05	0.18	0.19	0.00	0.19	1.02	0.00	0.00	0.01	0.00	0.00
Al	9.19	3.46	3.42	1.16	3.43	0.00	1.99	2.00	1.98	2.00	2.01
Cr	0.02	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe	1.58	2.01	2.00	0.00	2.92	0.95	2.38	2.15	1.72	2.14	2.48
Mn	0.04	0.02	0.01	0.00	0.01	0.00	0.07	0.26	0.49	0.26	0.07
Mg	0.52	2.66	2.69	0.00	1.92	0.00	0.21	0.11	0.07	0.11	0.22
Ca	0.00	0.00	0.00	0.16	0.00	0.00	0.39	0.54	0.81	0.54	0.27
Na	0.00	0.11	0.11	0.80	0.04	0.00	0.01	0.01	0.00	0.01	0.01
K	0.00	1.60	1.64	0.00	1.64	0.00	0.00	0.00	0.00	0.00	0.00
Total	15.37	15.49	15.52	4.97	15.54	1.98	8.03	8.04	8.04	8.04	8.02

defined by the biotite–chlorite–hornblende–plagioclase–garnet stability field (Fig. 6a). During this event, the garnet-in and then white mica-out boundaries had to be crossed to form the peak observed mineral assemblage. *P–T* conditions for the biotite–chlorite–hornblende–plagioclase–garnet stability field are between 565–700 °C and 3.0–9.5 kbar, and poorly constrain M2 conditions. To overcome this poor control on peak *P–T* conditions, contouring for composition isopleths of X_{Fe} in S2 biotite (0.50), amphibole (0.60) and unzoned garnet (0.84; Table 2), as well as $z(\text{Na})$ for hornblende ($\text{Na}[\text{M4 site}] = 0.14$; see Dale *et al.*, 2005) was performed. Relatively parallel X_{Fe} isopleths from biotite, amphibole and garnet, as well as pressure constraints using $z(\text{Na})$ in hornblende improved the estimate for peak *P–T* conditions to 6.3–6.6 kbar and 590–620 °C (Fig. 6a).

Proximal to HSZ – M3

Sample ST17. Evidence for only the S3 foliation is preserved in sample ST17. The metamorphic mineral assemblage that defines this foliation is composed of chlorite and biotite. Staurolite and garnet overgrow the foliation and define the peak metamorphic assemblage. Growth of these minerals requires the prograde *P–T* evolution of the sample to pass from a field of coexisting biotite–chlorite, into a garnet-bearing, and then staurolite-bearing fields (Fig. 6b). The growth of biotite prior to garnet constrains the lower *P–T* section of this path to where biotite is stable and garnet is not (Fig. 6b; < 3.5 kbar). The *P–T* evolution of the sample then passes up pressure and temperature into the staurolite-bearing field. Minimum temperature conditions associated with the peak preserved mineral assemblage are constrained by the limits of staurolite stability between 540 and 580 °C (Fig. 6b). The upper temperature limit of the stability field is constrained by the staurolite-out line. On the basis of the relatively

poor peak pressure estimates that are inferred from the large stability field presented in the pseudosection, contouring this field for composition isopleths of X_{Fe} and X_{Ca} for the unzoned M3 garnet (0.71 and 0.07 respectively) further constrains *P–T* estimates to 5.0–5.2 kbar and 600–640 °C (Fig. 6b).

Sample ST37. Sample ST37 from within a HSZ provides a useful comparison with the M3 *P–T* estimate constrained by the compositional isopleths from sample ST17. The pseudosection calculated from the bulk rock chemistry from sample ST37 is presented in Fig. 6(c). The S3 fabric within sample ST37 defines the main deformation-induced foliation, and is composed of fine-grained white mica, minor chlorite and biotite that is intergrown with a recrystallized polygonal quartz–plagioclase matrix. The stability field for this mineral assemblage is poorly constrained to the relatively large, high variance biotite–chlorite–muscovite–plagioclase–ilmenite–quartz field (Fig. 6c). The staurolite and biotite that overgrow the S3 foliation indicates the prograde evolution of the sample crossed into the mineral assemblage field of staurolite stability (Fig. 6c). The maximum temperature conditions for the staurolite–biotite–chlorite assemblage are primarily constrained by the garnet-in reaction line, as garnet is absent from the sample (Fig. 6c). The size of this stability field is substantially reduced because chlorite occurs within the peak mineral assemblage (Fig. 6d). This constrains *P–T* conditions for the M3 event to 4.0–5.0 kbar and 565–580 °C. This estimate is consistent with peak pressure conditions and lower temperature conditions calculated from the garnet compositional isopleths in sample ST17.

T–M_{H₂O} pseudosections

The role of fluid as a catalyst for metamorphic reactions is relatively well known (e.g. Fyfe *et al.*, 1978;

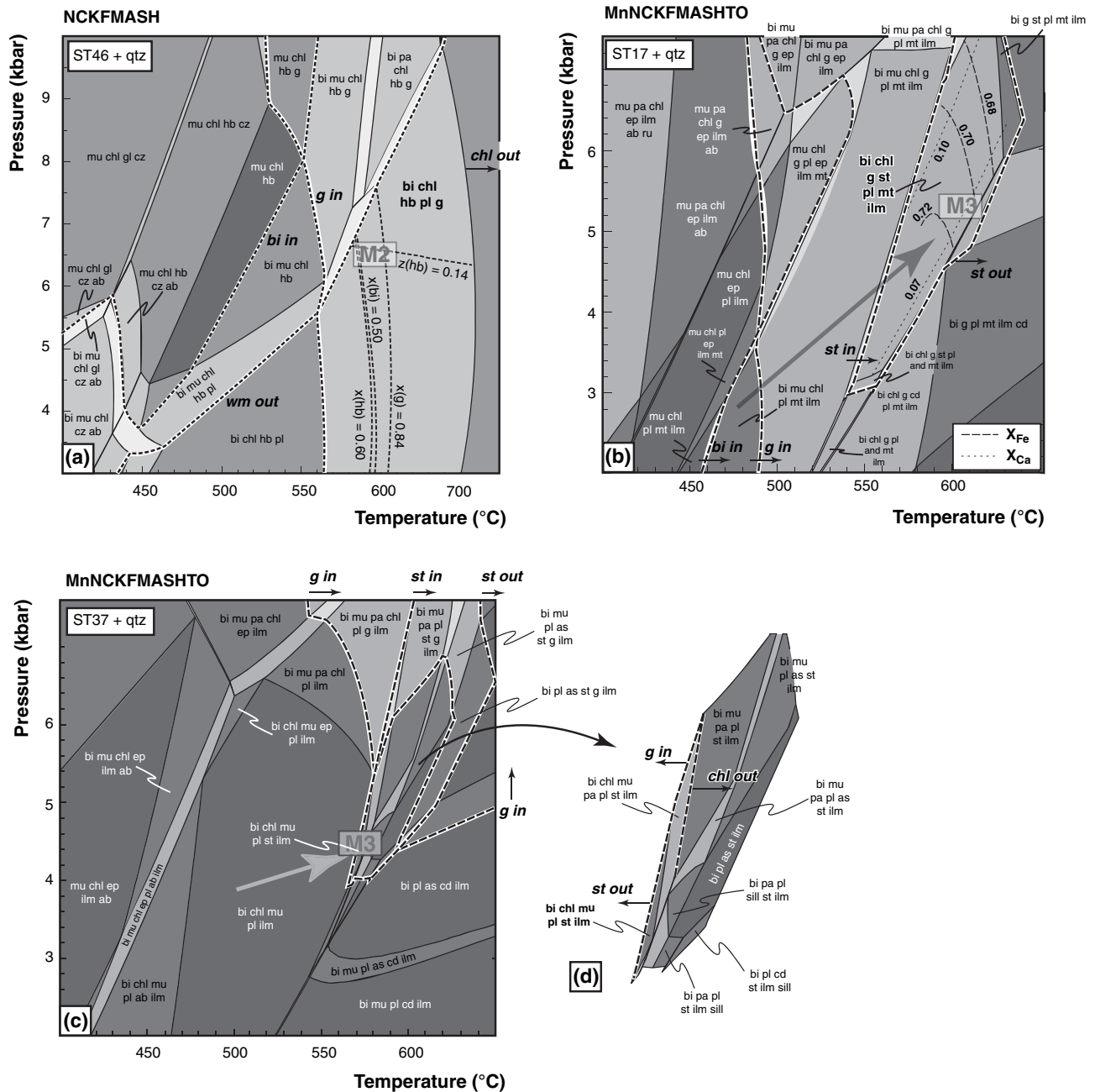


Fig. 6. Pseudosections in the systems NCKFMASH and MnNCKFMASHTO that model mineral equilibria for samples ST46, ST17 and ST37. The pseudosections were calculated using normalized bulk rock chemistry acquired by XRF analyses. H_2O and quartz are in excess. Variance of field relates to degree of shade; light shading – low variance; dark shading – high variance. (a) Pseudosection in the system NCKFMASH that models mineral equilibria for sample ST46. X_{Fe} composition isopleths for biotite, hornblende and garnet and X_{Na} composition isopleths for hornblende further constrain the peak P – T conditions. XRF acquired chemistry converted to mol. %: SiO_2 : 64.72; Al_2O_3 : 8.20; CaO : 3.24; FeO : 12.13; K_2O : 0.49; MgO : 8.27; Na_2O : 1.11. (b) Pseudosection in the chemical system MnNCKFMASHTO that model mineral equilibria for sample ST17. Arrow represents the M3 P – T evolution. X_{Fe} and X_{Ca} composition isopleths for garnet further constrain peak P – T conditions. XRF acquired chemistry normalized to mol. %: SiO_2 : 74.58; Al_2O_3 : 7.42; CaO : 1.34; FeO : 9.10; K_2O : 0.85; MgO : 4.89; Na_2O : 0.69; TiO_2 : 0.41; MnO : 0.29. Estimated value of Fe_2O_3 : 0.5. (c) Pseudosection in the chemical system MnNCKFMASHTO that model mineral equilibria for sample ST37. Arrow represents the M3 P – T evolution. XRF acquired chemistry normalized to mol. %: SiO_2 : 67.24; Al_2O_3 : 12.79; CaO : 0.70; FeO : 6.50; K_2O : 2.51; MgO : 7.19; Na_2O : 1.94; TiO_2 : 0.73; MnO : 0.06. Estimated value of Fe_2O_3 : 0.1. (d) Enlargement of stability fields that characterize the evolution of the peak mineral assemblage within sample ST37.

Thompson, 1983; Rubie, 1986; Guiraud *et al.*, 2001; Elmer *et al.*, 2006; Tenczer *et al.*, 2006). In this section, we examine the evolution of M_{H_2O} (as mole proportion of available H_2O ; Fig. 7a,b), through increasing temperature during M2 and M3 events. The T - M_{H_2O} pseudosections model mineral assemblage evolution with respect to bulk rock H_2O contents. This is done to elucidate the availability of H_2O fluid during the M2 and M3 events in the absence of infiltration of fluid from an external source.

Distal to HSZ – M2

Sample ST46. The relationship between temperature and M_{H_2O} in sample ST46 during the M2 event is constrained by the T - M_{H_2O} pseudosection presented in Fig. 7(a). The diagram represents the evolution of fluid (as mole proportion of H_2O) within the sample across a range of temperatures at constant pressure. Based on the peak M2 pressure conditions calculated for sample ST46, the T - M_{H_2O} diagram was constructed at 6.5 kbar.

The evolution of M_{H_2O} during M2 and is characterized by the progression of mineral equilibria across the biotite-in and then garnet-in lines (Fig. 7a). The path indicated by the arrow in Fig. 7(a) follows the threshold of H_2O saturation based on the assumption that small proportions of free fluid reside along grain boundaries (Guiraud *et al.*, 2001). During the growth of biotite and then garnet there was limited fluid production, which is indicated by the near-vertical H_2O saturation line. Fluid production within the sample changes at 560 °C where the H_2O saturation line becomes near-horizontal, coinciding with significant dehydration with heating as chlorite breaks down (Fig. 7a). The breakdown of chlorite results in a reduction in the amount of H_2O within the rock (from M_{H_2O} 0.32 to <0.14). The observed M2 mineral assemblage occurs at temperature and M_{H_2O} conditions between 590–660 °C and 0.22–0.12 respectively.

Proximal to HSZ – M3

Sample ST17. The evolution of temperature and M_{H_2O} for sample ST17 is presented in Fig. 7(b). Based on the peak pressure conditions calculated in the P - T pseudosection (Fig. 6b), the T - M_{H_2O} pseudosection was constructed at 5 kbar. The prograde evolution of this sample is highlighted by narrow intervals of pronounced dehydration (with respect to increasing temperature) that coincide with the breakdown of muscovite, and then chlorite (Fig. 7b). The growth of staurolite characterizes the peak preserved mineral assemblage, which also corresponds to a period of dehydration within the sample as muscovite breaks down (Fig. 7b). The peak preserved mineral assemblage for sample ST17, constrained by the temperature estimate of 610 °C coincides with the breakdown of chlorite, which would result in another

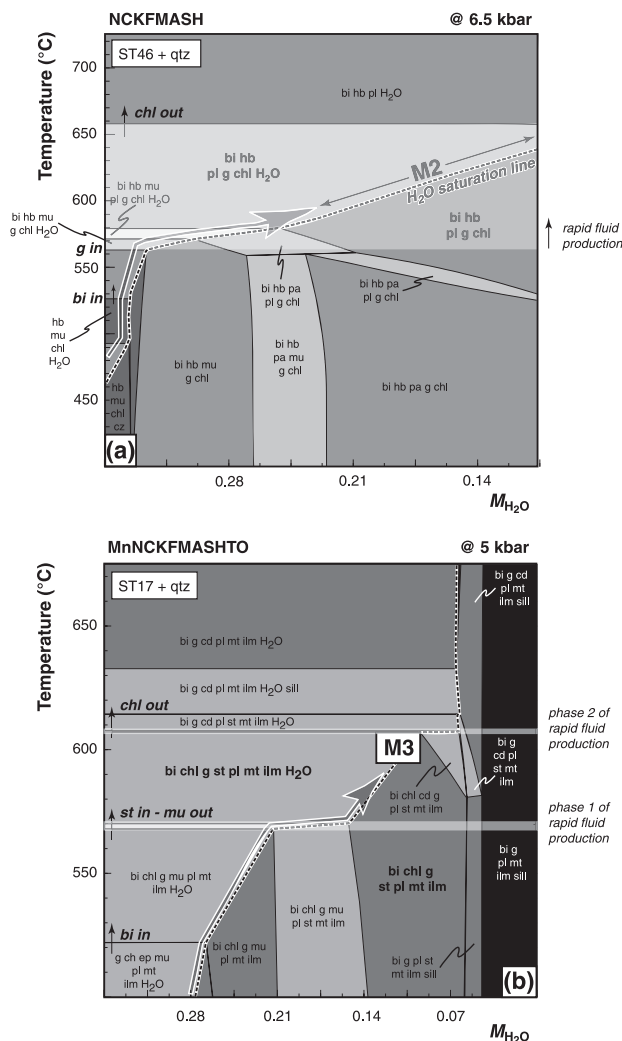
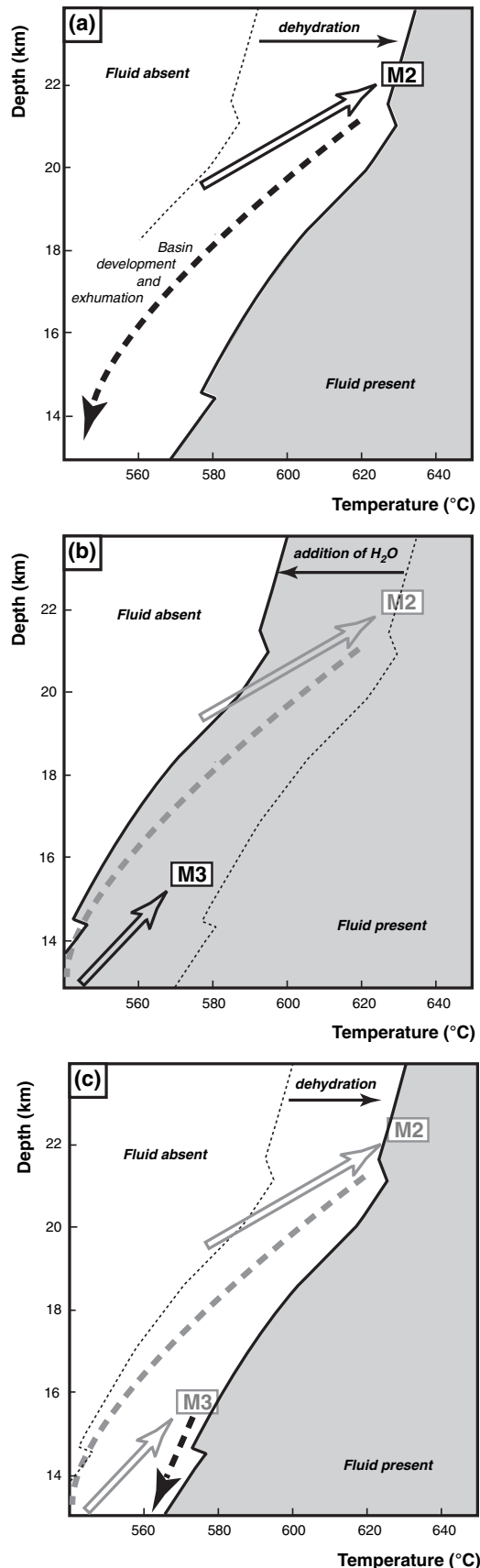


Fig. 7. T - M_{H_2O} pseudosections for samples ST46 and ST17. Thick dashed line represents H_2O saturation line. (a) T - M_{H_2O} path for sample ST46 follows the water saturation line until peak temperature (constrained by P - T pseudosection – Fig. 6a) is reached. This point represents the dehydration of H_2O in the sample and termination of metamorphic reactions. Based on the range of temperature constrained for the M2 event (Fig. 6a), a range of M_{H_2O} has also been assumed. The window of rapid dehydration as mineral equilibria crossed the chlorite-out line is shaded white. XRF acquired chemistry normalized to mol. %: SiO_2 : 64.72; Al_2O_3 : 8.20; CaO : 3.24; FeO : 12.13; K_2O : 0.49; MgO : 8.27; Na_2O : 1.11. Range of H_2O values: 15.00–0.00. (b) T - M_{H_2O} path for sample ST17. The windows of rapid dehydration as mineral equilibria crossed the muscovite- and chlorite-out lines are shaded white. The M3 peak temperature is constrained by the P - T pseudosection shown in Fig. 6(b–d). XRF acquired chemistry normalized to mol. %: SiO_2 : 74.58; Al_2O_3 : 7.42; CaO : 1.34; FeO : 9.10; K_2O : 0.85; MgO : 4.89; Na_2O : 0.69; TiO_2 : 0.41; MnO : 0.29; Fe_2O_3 : 0.5. Range of H_2O values: 15.00–0.00.

phase of pronounced fluid production and subsequent dehydration of the sample (Fig. 7b). The observed M3 mineral assemblage occurs at temperature and M_{H_2O} conditions between 590–610 °C and 0.14–0.10 respectively.



M3. Based on this evidence, fluid infiltration from an external source must have occurred throughout the Ruker Province during the M3 event.

DISCUSSION

Structural-metamorphic framework

The earliest fabric identified within the structural-metamorphic record is the bedding parallel S1 foliation that is preserved within the Tingey Complex. D2 folding resulted in the crenulation of this fabric and the development of the axial planar S2 fabric. The timing of the M2 event is unclear, yet can be constrained between the maximum depositional age of the Menzies and Stinear Groups at *c.* 2800 Ma and the maximum depositional age of the Sodruzhestvo Group at *c.* 950 Ma (Table 2). The *P-T* conditions during the M2 event are between 6.3–6.6 kbar and 590–620 °C (which equates to crustal depths of 22.5–23.6 km depth for an overburden density of 2.85 g cm⁻³) (Fig. 9a).

The *P-T* evolution within the HSZ during the M3 event is characterized by increasing pressure and temperature from 450–500 °C and <3.5 kbar (<12.5 km) to peak conditions between 4.0–5.2 kbar (crustal depths of 14.3–18.6 km) and 565–640 °C (Fig. 9b). The *P-T* path defining the evolution of sample ST17 starts at lower pressure conditions than attained during peak M3 metamorphism (Fig. 6b). This suggests that between the M2 and M3 event exhumation of the Tingey Complex to crustal depths less than ~12.5 km occurred. This is consistent with basin development that resulted in the deposition of the Sodruzhestvo Group, prior to the M3 event (Fig. 9b).

Controlling factors of early Palaeozoic M3 metamorphism

Fluid content

The metamorphic mineral assemblages preserved in the HSZ rocks of the Ruker Province are consistent with fluid influx during the M3 event. Dehydration of the Ruker Province during the M2 event would have

Fig. 9. Summary depth–*T*–*M*_{H₂O} diagrams of the metamorphic and H₂O evolution of the Tingey Complex across the M2 and M3 events. H₂O saturation line topology is from the pseudo-section shown in Fig. 8. (a) Dehydration during the M2 event resulted in the H₂O saturation line moving up temperature. The post-M2 tectonic evolution of the Tingey Complex occurred within a dehydrated environment, and therefore, mineral equilibria were not preserved during changes in *P-T* (thick dashed line). (b) Addition of fluid during the M3 event resulted in the H₂O saturation line moving down temperature. This was required for the M3 mineral equilibria to proceed (represented by black open arrow – from M3 *P-T* paths in Fig. 6b,c). (c) Dehydration during the M3 event resulted in the H₂O saturation line moving up temperature. The retrograde path (thick dashed black line) of the M3 event would not be preserved by mineral equilibria because it was located within the fluid absent field.

resulted in fluid-absent conditions for M3 (Fig. 9a). Therefore, preservation of petrographic evidence for the exhumation of the Tingey Complex between the M2 and M3 would be unlikely (dashed line – Fig. 9a). The location of the M3 P – T conditions within the H_2O -absent field requires open system behaviour for metamorphic reactions that consume H_2O to proceed. The addition of fluid means the H_2O -in line shifts to lower temperature, which subsequently allows the M3 metamorphic reactions to proceed (Fig. 9b). Following this, dehydration as a result of increasing temperature during the M3 event explains the lack of retrograde features associated with the M3 mineral assemblages within the HSZ rocks (Fig. 9c). This may be consistent with the fluid influx being relatively short lived, and not outlasting deformation.

Deformation

Samples ST17, ST46 and ST50, which are located away from the HSZ, preserve S1 and S2 foliations as well as M3 retrograde assemblages. Conversely, samples ST17 and ST37, which are located within the D3 HSZ, preserve evidence for only the S3 foliation. We infer that thorough recrystallization and development of the M3 mineral assemblages was related to the intensity of strain and the influx of fluid through the more permeable, sheared rocks within the D3 HSZ. The P – T – M_{H_2O} diagrams shown in Fig. 9(a–c) indicate that the open system behaviour is required to explain mineral equilibria preserved within the Ruker Province. The cessation of deformation along these shear zones was probably associated with the termination of fluid saturation, indicated by the lack of M3 retrograde assemblages within the HSZ rocks. Conversely, less hydration occurred within the lower permeable non-sheared rocks located away from the HSZ (i.e. samples ST19 and ST50). This resulted in partial recrystallization of the M1 and M2 mineral assemblages during the M3 event.

Tectonic implications

Tectonics and fluid flow

Modelling the evolution of available fluid H_2O between the M2 and M3 events indicates that hydration of the middle crust must have occurred during the early Palaeozoic M3 event. A mechanism that can explain fluid transport into the middle crust is therefore required. Jessell & Lister (1991) demonstrated that increasing strain rates are responsible for the localization of deformation, particularly between two different rock types. The identification of the HSZ between the lithological contacts of the Tingey Complex (Fig. 2a) supports this. Evidence for the coupling of increasing strain with fluid flow (Oliver, 1996) provides a mechanism for the transport of fluid, which would have resulted in the hydration of the HSZ rocks

due to dilatancy pumping (Nur & Brooker, 1972). This fluid transport system would not have developed pervasive flow throughout the terrane. Instead, increased flow would be promoted within the more dilate HSZ. Limited hydration of rocks distal to the HSZ relates to lower levels of permeability within the non-sheared rock units.

Without stable isotope data, determining the origin of the fluid is a difficult task. The amount of fluid H_2O in the crust can be influenced by the input of fluid from a non-lithospheric source (i.e. meteoric), the release of fluid from minerals during metamorphism and/or the release of fluid from a crystallizing magma. The input of fluid into the middle crust (which is required in this study) could have occurred from the surface, requiring a gravity-driven or flow-controlled compressive tectonic setting (Sibson *et al.*, 1975; Connolly & Podladchikov, 2004; Mulch *et al.*, 2004) or from below, such as during subduction or lower crustal metamorphism (Fyfe *et al.*, 1978; Bebout & Barton, 1993). The potential fluid source could be determined through studying the stable isotope record of samples distal and proximal to the HSZ to determine the nature of the fluids (i.e. meteoric, crustal or mantle derived). Without these data, possible mechanisms that may have resulted in the hydration of the middle crust are discussed herein.

In support of a meteoric source, it has been argued that regional extension can provide a mechanism to overcome hydraulic head for surface waters to penetrate the middle crust (Frike *et al.*, 1992). In addition, compressive tectonic settings also provide a mechanism by which upper crustal fluids could be swept downward into the middle crust (Connolly & Podladchikov, 2004). Basin development during the middle Neoproterozoic, which was followed by inversion during the M3 event (Phillips *et al.*, 2006, 2007), provides one possibility for the attainment of meteoric fluid within the middle crust during M3 metamorphism.

Alternatively, if the source of fluid was the lower crust, a model of subduction, or deep crustal metamorphism is required. Advection of fluids during dehydration melting of subducted oceanic or accretionary wedge material would result in hydration of the overlying lithosphere. This model requires an active convergent setting for the Ruker Province during the early Palaeozoic, as well as an infinite source of fluid – which is consistent with ongoing subduction.

Deep crustal metamorphism could also provide a fluid source for mid-crustal hydration. In the case of this study, upper amphibolite facies metamorphism could have occurred within a lower crustal position (below the currently exposed crustal level) during the M3 event (Elmer *et al.*, 2006). Fluid released during lower-middle crustal metamorphism would infiltrate the overlying rocks now exposed within the Ruker Province. This model assumes a finite fluid source as the lower to middle crustal rocks undergo dehydration.

Implications for current tectonic models

The implications of the new constraints on fluid flow for the conflicting early Palaeozoic tectonic models for the Ruker Province are discussed herein. Two contrasting tectonic models that explain the origin of early Palaeozoic orogenesis within the Ruker Province have been presented in the literature. One model locates a collision between the Rayner Complex (and attached India) and the Ruker Province that occurred during the assembly of Gondwana (Fig. 10a; Boger *et al.*, 2001). The alternate model suggests that early Palaeozoic tectonism was caused by the reactivation and inversion of Neoproterozoic intra-cratonic basins (Phillips *et al.*, 2007). Inversion was driven by far-field stresses created during the collision between East and West Gondwana along the Mozambique suture (Fig. 10a; Phillips *et al.*, 2007). The two models require distinctly different tectonic settings for the Ruker Province during the early Palaeozoic.

For the collisional model, dehydration of a subducting slab below the Ruker Province provides a source of fluid that would be transported into the middle crust during the development of the HSZ (Fig. 10b). However, there is evidence that dehydration of the HSZ rocks occurred during prograde metamorphism, as indicated by the lack of M3 retrograde assemblages preserved within these zones. Dehydration of the HSZ during prograde metamorphism can be interpreted in two ways: termination of deformation (i.e. dilatancy pumping along the HSZ ceases) or termination of the fluid source (i.e. subduction ceases). The former interpretation is unlikely as the M3 prograde paths determined in this study would relate to crustal thickening caused by shortening. During this phase of orogenesis, fluid flow should be uninhibited as the high intensity of strain within the HSZ would be proportional to the rate of fluid transport (Oliver, 1996). Therefore, for the subduction model to accurately explain the evolution of lithospheric H₂O during the early Palaeozoic the latter must be assumed – dehydration was due to the cessation of subduction.

In contrast to the subduction setting, we suggest that a tectonic model that evokes a limited period of fluid influx is more realistic in describing the early Palaeozoic evolution of lithospheric H₂O. The development of basins and deposition of the siliclastic dominated Sodrzhestvo Group between 950–550 Ma (Phillips *et al.*, 2006) occurred between the M2 and M3 events. Basin development probably resulted in meteoric fluid flow along extensional detachments within the upper crust as well as lithospheric thinning and metamorphism within the lower crust (Sandiford & Powell, 1986; Frike *et al.*, 1992) (Fig. 10c). The exhumation of the Tingey Complex into the upper 10–15 km of the crust prior to the M3 event, as suggested through mineral equilibria modelling of the M2 and M3 events, can be explained by this period of extension. Upper

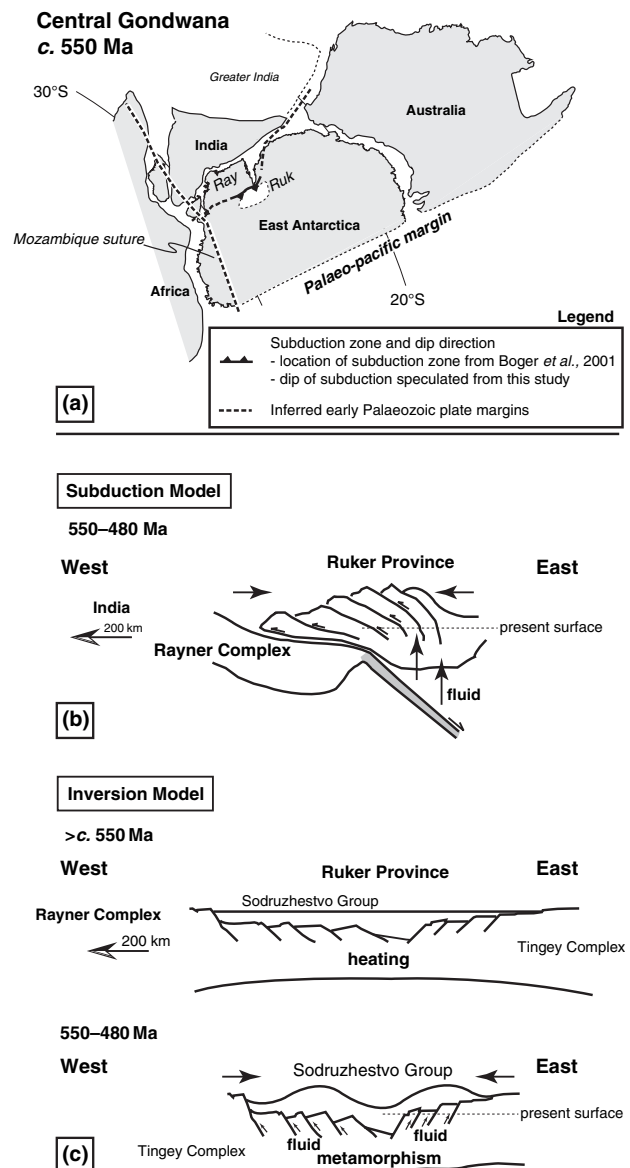


Fig. 10. Map of East Gondwana during the early Palaeozoic, and two tectonic scenarios that can explain middle crustal hydration within the Ruker Province during the M3 event. (a) Map of East Gondwana, and the location of inferred plate boundaries during the early Palaeozoic. The dip of the subduction zone within the Ruker Province is based on fluid flow constraints from this study. Palaeolatitudes are from Meert (2003) and Collins & Pisarevsky (2005). Abbreviations: Ray – Rayner Complex; Ruk – Ruker Province. (b) Subduction model – dehydration melting of the down-going slab releases fluid that is transported into the middle crust. (c) Lower-middle crustal heating resulting from Neoproterozoic basin development. Metamorphism of the lower-middle crust during basin inversion releases fluid that is transported into the middle-upper crust by dilatancy pumping along the basement HSZ.

amphibolite metamorphism during crustal thickening associated with inversion of the basin (M3 event) resulted in dehydration of the lower crustal rocks. In

addition, middle to lower crustal metamorphism during this time may have been accentuated by incubation by the thick sedimentary blanket representing the Neoproterozoic Sodruzhestvo cover succession (Sandiford *et al.*, 1998). These processes provide a mechanism for a finite lower crustal fluid source, which infiltrated the middle crust during dilatancy pumping along the HSZ. Furthermore, extension-related detachments within the upper crust could have resulted in gravity-driven flow of meteoric water from the surface. Evidence for elevated levels of upper crustal hydration (cf. middle crustal levels) is the consistent retrogression of M3 assemblages (i.e. chlorite pseudomorphs of garnet – Fig. 5h) within the upper stratigraphic levels of the Sodruzhestvo Group.

CONCLUSIONS

The metamorphic framework presented in this study constrains two key metamorphic events that affected the Ruker Province of East Antarctica. The M2 event is constrained to the Proterozoic and occurred at amphibolite facies P – T conditions. The event is associated with the regional folding of the Tingey Complex into kilometre scale upright to reclined folds. M3 metamorphism affected the Ruker Province during early Palaeozoic times and is associated with a dominant phase of fluid influx into the middle crust. Sources for this fluid could be the surface and lower crust, represented by meteoric and fluid generated during metamorphism within a lower crustal position respectively. A model of Neoproterozoic basin development best explains these two sources. Fluid transport during early Palaeozoic convergence was along HSZ, which resulted in the localization of the M3 mineral assemblages proximal to these structural zones. The preservation of the M3 mineral assemblages was highly structurally dependent, with thorough recrystallization of rocks during the M3 event within the hydrated and intensely sheared HSZ. The lack of retrograde mineral assemblages within these rocks suggests dehydration occurred during the prograde path. We interpret this to reflect a finite fluid source from the lower crust, which favours deep crustal metamorphism, rather than the subduction model.

ACKNOWLEDGEMENTS

The Australian Antarctic Division (AAS Grant 1215) and the BGR (Bundesanstalt für Geowissenschaften und Rohstoffe) are thanked for logistical and financial support for the 2002–03 PCMEGA field season in Antarctica. G.P. acknowledges the support of a Bargwanath and Albert Shimmins Scholarship held at the University of Melbourne. R.W.W. acknowledges support from Australian Research Council grant DP 0557013 and a Melbourne University Research grant scheme grant. Additional help with THERMOCALC from

R. Powell is appreciated. K. Stüwe, G. Rebay and D. Tinkham are thanked for insightful comments that improved the manuscript.

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Received 18 December 2006; revision accepted 30 June 2007.